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GLACIOLOGY, unlike glacial geology, is concerned primarily with the regimen of "living" glaciers. Although glaciology dates back to the classical observations of Agassiz on the glaciers of the Alps and has received intermittent attention of various workers since, it is only during the last decade that it has attained the status of an independent earth science. In view of the important relations between glaciology and both glacial geology and climatology, we are devoting this issue of the *Journal* to a group of papers on glaciology.

Among the papers to appear in forthcoming issues of the *Journal* are:

"Submarine Canyons Bordering Central and Southern California." By JOHN C. CROWELL

"Origin and Classification of Chemical Sediments in Terms of pH and Oxidation-Reduction Potentials." By W. C. KRUMBEIN and R. M. GARRELS

"Surface Area of Deep-Sea Sediments." By J. LAURENCE KULP and DONALD R. CARR

"Evaluation of Criteria for the Forcible Intrusion of Magma." By JAMES A. NOBLE

"Use of Soil Mechanics Data in Correlation and Interpretation of Lake Agassiz Sediments." By J. F. ROMINGER and P. C. RUTLEDGE

"Composite Origin of Submarine Canyons." By FRANCIS P. SHEPARD

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SMITH, A. V. (1926) Discovery of fish remains in Ordovician rocks of the Black Hills (abstr.): Geol. Soc. America Bull. 52, p. 27.
——— and DUNSANY, A. J. (1929*a*) Early Ordovician faunas of South Dakota: U.S. Geol. Survey Bull. 444, pp. 1–76.
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INTRODUCTION TO ICE PETROFABRICS[1]

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ABSTRACT

The mechanical properties of ice, particularly viscosity, depend on its structure and texture. The study of ice petrofabrics is an essential auxiliary to all experimental work on ice mechanics and is of value in the study of glacier flow. A common interlocking type of glacier ice structure is described in detail. Crystal orientation patterns of dead and living ice are illustrated. There is need for field and laboratory work to correlate stress-strain and crystal-orientation patterns. This paper is mainly a description of ice petrofabric techniques for field and laboratory use.

INTRODUCTION

The mechanical properties of ice are not well known, although there is a large literature on the subject. Most laboratory work was done under insufficiently well-controlled conditions to be of more than general interest. From the point of view of glacier flow, the most important property of ice is its viscosity, published values of which are useless as a basis for mathematical treatment of glacier flow. It is now known that viscosity of ice is a function of at least the following variables:

1. *Temperature.*—Viscosity decreases as temperature increases.

2. *Stress.*—Ice as a fluid is non-Newtonian, the velocity of deformation increasing faster than the load. In rheological terms, wet ice is almost certainly visco-elastic, whereas at lower temperatures it may be plasto-elastic, i.e., there may be a threshold stress below which the strain is purely elastic.

3. *Grain size.*—Fine-grained ice is much less viscous than coarse-grained ice.

4. *Grain shape.*—Ice with granular, equidimensional crystals is less viscous than if the grains are interlocking.

5. *Grain orientation.*—Preferred orientation makes the ice body significantly anisotropic in respect to viscosity. The crystallographic basal plane is the only reported plane of lattice slippage. Whether any polarity of slippage exists within the hexagonal or trigonal symmetry within this plane is not known.

The grain properties influencing viscosity have not been given sufficient consideration in theories of glacier flow. Preliminary laboratory experiments, for example, indicate that fine-grained névé ice is less viscous than coarse-grained ice by several orders of magnitude. A long period of patient petrographic and petrofabric work on glacier ice in the field and

[1] Manuscript received February 26, 1951.

[2] On leave of absence from Rutgers University.

on glacier and artificial ice in the laboratory as an essential auxiliary to work on ice mechanics must elapse before a good theory of glacier flow can be formulated.

Field work done during the past decade, together with older observations thinly scattered throughout glaciological literature, show that preferred orientation is the rule in glacier ice and random orientation the exception.

Since many different orientation patterns have been recorded, it is reasonable to assume that the ice aggregate structurally reflects the pattern of stress and strain. Ice petrofabrics, then, can become a most valuable key to the estimation of stress distribution in glaciers, but the lock has yet to be constructed. The relation of crystal orientation to stress-strain pattern must be determined by laboratory and field work, as well as the time lapse between a change of stress or strain, either in magnitude or in direction, and the change of orientation pattern. Experiments by Tarr and Von Engeln (1914) indicate that under high stress the time lapse can sometimes be very short. This time lapse with the persistence of relic structures is important, as it may, on the one hand, yield information on past stress-strain history of a specimen and, on the other hand, it is likely to render difficult the interpretation of a structure. The late Max Demorest had begun systematic laboratory work on the relation between stress-strain and change of texture and orientation pattern, but did not publish his results. There is some preliminary work of the author (1938), and orientation diagrams of névé and glacier ice were made by Perutz (Perutz and Seligman, 1939).

Petrofabrics is the detailed study of structure and texture of rocks. In ice the composition is a simple one, comprising, at most, four phases: ice crystal and water, air, and mineral inclusions. The description of an ice aggregate consists mainly in the description of its geometry —the shape, orientation, and spatial relationships of its elements.

In the following section some techniques of ice petrofabrics are described, together with examples of results. These techniques must be further developed and new ones added.

The purpose of this paper is to stimulate a wider interest in ice petrofabrics. The subject is worthy of major attention, for, until it is well advanced, physical glaciology will remain without a solid foundation. Above all, no laboratory work on ice mechanics should be done without careful petrofabric study of the specimens to be tested. A cursory inspection of sections between crossed polaroids is inadequate.

TECHNIQUES FOR OBTAINING SPECIMENS FROM GLACIERS

The ice specimens for petrofabric and other studies can be obtained from glaciers in two ways—they are excavated by hand from accessible places or are brought up from greater depth by a coring device.

Unless the "weathering" of ice is to be studied, it is advisable to work only on ice that has not been exposed to the corrosive effects of sunlight. Ice exposed only to sky radiation or ice water is usually perfectly sound. Specimens should therefore be obtained from under water, from under moraine cover, or from permanently shaded crevasses or moulin holes, etc. Considerable labor (1–3 hours) is necessary for carving an oriented squared block of approximately 1 cubic foot. It is a matter of slow and patient excavation and careful chipping. A hurried job usually results in a cracked block of doubtful orientation. It is convenient

to bevel one corner of the block for an orientation reference point. The use of a large-diameter, hand-operated coring cutter some 2 feet long would doubtless save much time in procuring large specimens and would permit more accurate orientation. The design of such a tool is available from the New England Division, Corps of Engineers, U.S. Army, Boston, Massachusetts.

Ice-carving tools are the ice pick, the chipping hammer, and the handsaw. The mountaineer's ice pick is quite useless, being too light and too blunt. Sharpness is the prime requisite of all ice-cutting tools, making a sharpening file an essential part of the equipment. The ice pick must have not only a sharp point but also the correct shape; too large or too small curvature renders a pick very inefficient. A pick copied from the model developed by Fritz Steuri, of Grindelwald, was found to be excellent, whereas another of slightly different curvature performed rather poorly. Figure 1 is a sketch of a Steuri-type excavating pick. The chipping hammer weighs approximately 1 pound and has a very sharp cutting edge at both extremities of the head. One edge is parallel to the handle, the other at right angles to it, permitting chipping from the side and from the front.

Fine-toothed saws are useless for sawing wet ice. Not more than nine teeth per inch is recommended, and a stiff saw is better than a flexible one. A rather heavy tooth offset is preferable. Even so, the teeth clog up with ice, making it necessary to pull out the saw after a few strokes and to slap it against a flat surface in order to clear the teeth. The sharper the teeth, the faster the cutting.

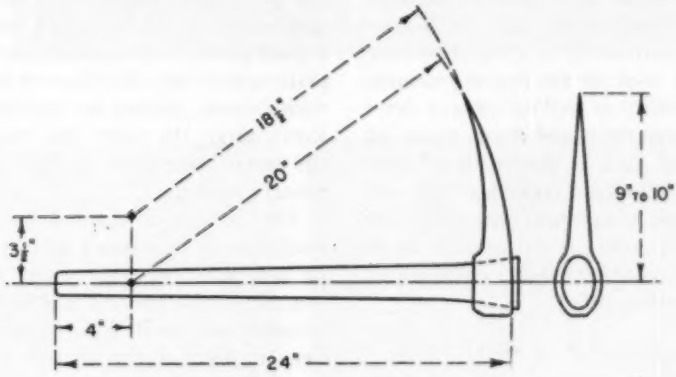


FIG. 1.—Steuri-type ice-excavating pick

Plate 1, *A*, shows some ice-cutting tools —chipping hammer, saw, and pick (needs sharpening).

Specimens of ice from a depth of several tens of feet can be obtained from deeper crevasses or potholes, but collection is dangerous. Deep core drilling is the only means of sampling to greater depth. Mechanical drilling with slightly modified standard diamond-drilling equipment was attempted with some success by the Juneau Ice Field Research Project of the American Geographical Society in 1950. Drilling technique will be described elsewhere.

Deep mechanical core drilling is unlikely to become popular in the near fu-

ture, as it requires very heavy equipment and is correspondingly expensive. It is, however, the only means of obtaining deep samples from "cold" high polar glaciers, where compressed air will have to replace water to remove the cuttings. For "warm" glaciers (ice at melting temperature throughout) an interesting alternative is being investigated, which may prove to be more useful and cheaper. This is a modification of the electric hot point. The heating element would be an annulus at the bottom of a light coring barrel. In order to be able to use light cable, a transformer could be lowered with the instrument or a high-resistance conductor used for the heating element. The possibility of pulling cores of determined azimuth would be a great advantage of such a device over rotary drilling. Cylindrical cores are very convenient for the study of consecutive thin-sections. A groove is cut parallel to the cylinder axis to establish a common azimuth reference point.

PREPARATION OF SECTIONS

Thin-sections are easily cut with a sharp saw. Larger sections can be cut by hand. It takes about ½ hour to make a 1-square-foot cut. Sections of wet ice should be cut not less than ½ inch thick, to prevent breakage. They can then be melted down to the desired thickness and size with a warm plate. A 12-inch aluminum frying pan full of warm water is a most satisfying tool for this purpose. The section is laid on paper to prevent slipping (a large supply of absorbent paper towels is almost indispensable for work with wet ice; also several pairs of cotton work gloves) and melted down first on one side and then on the other, by a circulatory motion of the warm plate to expel the meltwater and yield a flat surface. Small rock fragments embedded in the ice are immediately detected by feel and must be removed with a sharp steel point.

At temperatures below the melting point, sections can be cut thinner to begin with, if a prepared surface of the specimen to be sawed is first fastened to a glass plate. A moderately warm glass plate melts a very thin layer of ice, which then freezes, making an excellent bond. Fortunately, the water film freezes with the same orientation as the crystal in contact with it.

The section is ground to required thickness on sandpaper and is polished on newsprint or rouge paper. Grinding is a rather slow process, unless the paper is cooled with solid carbon dioxide to prevent melting of the section. If optical anomalies of strained crystals are to be studied, the specimen should be prepared cold, as relaxation is rapid at or very near the melting point.

The solid section, if thin, must be protected from evaporation by covering with glass. Optical contact is established by an inert liquid or by partial melting of the section by the warm cover glass.

For general work with cold ice the ambient temperature should not exceed

PLATE 1

A, Some tools for ice work: excavating pick, chipping hammer, handsaw, steel point, hot plate (frying pan and blowtorch), microscope, frame for preparation of consecutive polished sections.

B, Three-axis universal stage for 9-inch square sections.

C, Immersed three-axis stage for 2.3-inch circular sections. Small knob at left rotates immersed section carrier. Azimuth is read on transparent circular scale under section carrier. Double-walled (cooling jacket) container rotates on ball bearings between 3-inch polaroids.

C

A

B

Equipment for preparation and study of ice sections

A

B

Dead and living ice of Malaspina Glacier

—4° C. Here a warning is indicated. In working with ice in a cold chamber (maximum air temperature −4° C. if the ice is to remain dry), inadequate clothing is a most serious health hazard, owing to lack of exercise. The habitual coldroom worker should *never* disregard two main precautions: keep the back (particularly the kidney region) always comfortably warm, and leave the cold room when a chill reaches above knees or elbows.

GRAIN SIZE AND GRAIN SHAPE

A "grain" of glacier ice (*Gletscherkorn*) is a single crystal, in contrast to snow, in which it may consist of several crystals more firmly cemented to one another than to the crystals of other grains. Thus, in glacier ice, grain size and shape are synonymous with crystal size and shape. One of the many striking properties of ice is its purity. The ice lattice appears to have an extremely small tolerance for molecular impurities; gases and other solutes in water apparently do not enter the crystal lattice in molecular dispersion. They are either excluded by the growing crystal or included in the form of bubbles (gas) or globules (liquid). Solid impurities also tend to accumulate at the crystal boundaries. The few parts per million of soluble impurities precipitated with snow and rain are thus concentrated in the thin intercrystalline layers of glacier ice. A block of clean glacier ice exposed to solar radiation will melt along all its grain boundaries, and the first meltwaters show higher electrical conductivity than subsequent samples do, thus proving a higher concentration of solubles at the grain boundaries (Renaud, 1949).

The intercrystalline boundary is presumably in a disarranged (noncrystalline) state, which may also account for preferential melting. Experiments with ice made from very pure water would be desirable. Whatever the reason—relatively high concentration of impurities or disarrangement—polycrystalline ice, etched by evaporation below the melting point or by slow surficial melting, reveals grain boundaries in the form of grooves. These grooves are approximately semicircular in cross section and make possible an imprint of grain boundaries on a flat ice surface by the technique of rubbings (Seligman, 1949) or castings (Schaefer, 1950). Castings cannot be made from wet ice. I obtained very good rubbings on stiff tracing paper (Post No. 173) with red grease pencil (Fisher "non-run"). It is advisable to use a relatively waterproof paper, so that the upper surface of the paper will not become wet before the rubbing can be finished. On looking at the reflection of light from a wiped, newly prepared surface at a flat angle, the grain boundaries appear as very fine lines; but in the ice laboratory 1½ hours' waiting was necessary for the grooves to become wide and deep enough to show clearly on rubbings.

In order to obtain a satisfactory picture of the structure of the coarse-grained dead ice fringing the Malaspina, a 9-inch cube was shaped with saw or chipping hammer and set in an aluminum

PLATE 2

A, Looking north over dead ice of Malaspina Glacier. Bare ice shows in two places (*dark*) in middle of picture. Whole landscape is underlain by ice, with no more than several feet of moraine cover. Trees are up to 90 years old.

B, Looking southwest from edge of living ice of Malaspina Glacier. Dark dead ice in background. Left foreground is clear ice, right foreground shows white bubbly ice with thin clear bands.

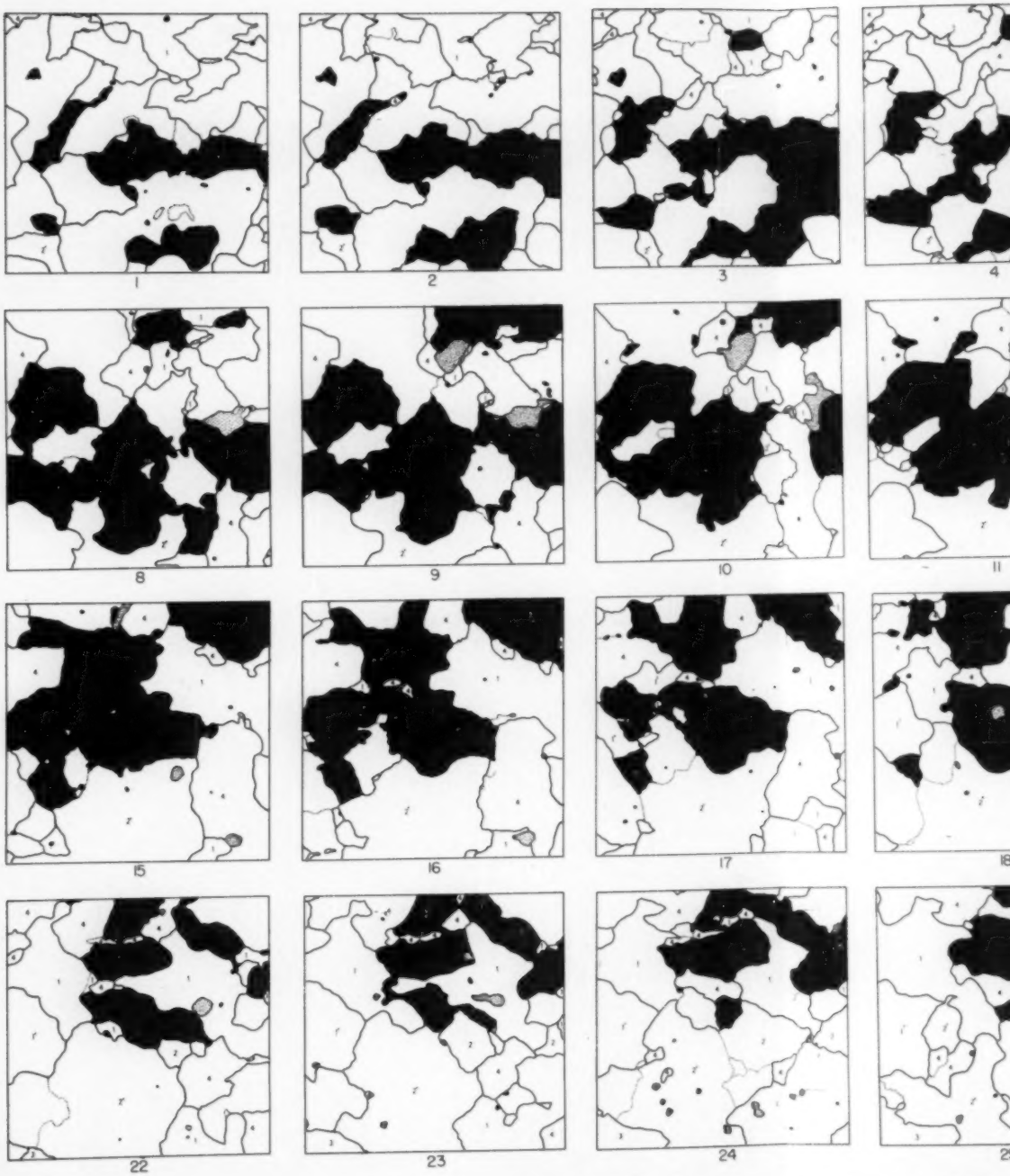
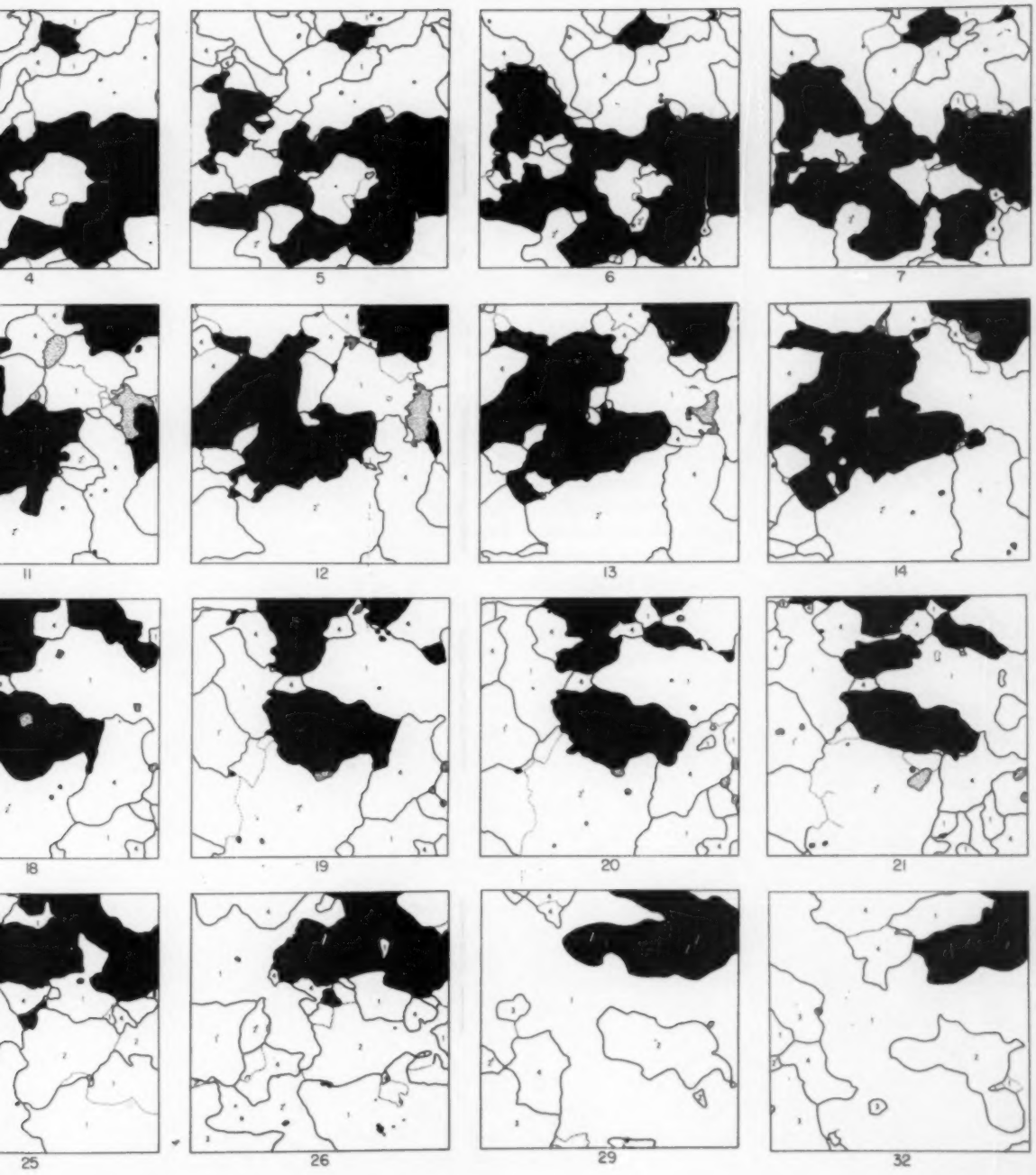


FIG. 2.—Dead ice from laboratory excavation. Consecutive polished se



ed sections (rubbings) at 0.2-inch intervals. Frames are 8 inches square

frame (pl. 1, *A*). In order to minimize melting at the bottom of the block and to prevent the block from slipping, a 7-inch-square array of nails was hammered into a plywood base and the heads snipped off, leaving ½ inch of the nails protruding. A 9-inch square of brass mosquito screen was pressed down over the nails onto the wood, and the ice block set on the nails. In a few minutes the nails had penetrated into the ice, holding the block immovable in reference to the frame.

The top part of the frame has an opening fitting over the ice block and can be lowered in relation to the block. The paper for the rubbings is pinned to holes in the frame top with nails, providing reference points on the paper. The following technique was used: the top of the frame was lowered until the ice block projected a few tenths of an inch. A 12-inch aluminum frying pan containing warm water was used to melt down the block to the level of the frame, leaving a plane horizontal surface. Whenever the pan scraped against even a very small sand grain enclosed in the ice, it had to be lifted off and the grain removed from the ice with a sharp steel point. The surface was finally polished with the edge of the hand. A pause of 1½ hours followed, to allow the grain boundaries to form grooves of sufficient size for rubbing. During this time the frame was covered with a sheet to slow the melting process. It was observed that grooves develop much better on slow than on fast melting. The scant meltwater was then wiped off with soft tissue paper and the rubbing made. The frame top was lowered 0.2 inch and the process repeated.

In this manner 26 rubbings were made, representing consecutive sections at 0.2-inch intervals over 5 inches of ice. Even in such coarse-grained ice with grain dimensions more than fifty times larger than the intervals between sections, it was sometimes difficult to correlate the grain boundaries of successive sections, owing to extremely irregular grain shape. Tracings of the rubbings are reproduced in figure 2, nos. *1–26* (bottom to top of the figures corresponds to a direction E. 70° N in the ice block). The frames are 8-inch squares. The dotted areas represent sections of inclusions of debris, rock fragments, and, more commonly, spherical pockets of very fine silt. The dotted areas are considerably larger than the debris cross sections, owing to melting of the surrounding ice on preparation. The full lines are grain boundaries. The dotted lines are doubtful grain boundaries, which are fuzzy on the rubbing, not due to faulty rubbing but to shallower and less distinct grooves. It is noticeable that these lesser grooves sometimes end in the middle of a crystal. The rubbings, especially those made after the overnight pause of some 8 hours, show a wealth of line detail, a discussion of which, at this stage of the investigation, would be premature. Careful study of good rubbings or castings or, better still, study of line detail on the ice surface itself will yield much information on the reasons for the growth of some crystals at the expense of others. Apparently, some of the less distinct grooves show old grain boundaries, and some may reflect stress patterns. The character of a groove may also depend on the relative orientation of the two crystals it separates.

In order to illustrate the most intricate shape of the majority of the few crystals (perhaps less than a dozen) in the 8×8×5-inch piece of ice, the most striking one (No. 5) has been blackened in the tracings. Thus the black areas, a series of sections of a single ice crystal, reveal a truly amazing shape. Some of the

other crystals are indicated by numbers. The orientations of these crystals were measured in three consecutive thin-sections (fig. 2, nos. *26*, *29*, *32*). No. *29* is $\frac{3}{5}$ inch above No. *26*, and No. *32* is the same distance above *29* (the ice block was laid in the frame upside down). No optical anomalies were observed, the different areas of a single crystal extinguishing equally and simultaneously.

Crystals Nos. *1* and *1'* are optically parallel, yet obviously two distinct crystals down to section No. *26*. But in section No. *29* the boundary line has disappeared. Crystals Nos. *2* and *2'* are also optically parallel, with weakly marked boundaries in sections *24* and *26*. One is tempted to conclude that the secondary crystallographic axes may rotate along the main axis. Evidence of this could perhaps be obtained by producing Tyndall melt figures (by focusing of radiant heat) in different parts of a single crystal.

It is observed that other crystals, such as Nos. *1* and *4*, have the same complicated shapes as No. *5* (black). The whole aggregate is an interlocked network of crystals. This was most strikingly demonstrated in the following manner: a large ice block (3 cu. ft.) was exposed to the sun. After several hours all the grain boundaries were completely melted out, the block being held together only by interlocking of the grains. Not a single grain could be extracted from the block without breaking it. A similar phenomenon was observed by Buchanan (1901) on ice from bergs floating on the Maerjelensee at the fringe of the Aletsch Glacier in the Alps; also by D. B. Lawrence (private communication) on icebergs in Glacier Bay, Alaska.

This type of interlocked structure appears to be quite common. In clear ice it can be recognized by the involved shapes of the often visible lines separating three grains. They look like a mass of tangled wire. I have observed this structure in the following materials: old dead ice, Malaspina Glacier; old living clear ice, Malaspina Glacier; old living bubbly ice, Malaspina Glacier; dead ice, Mendenhall Glacier; living clear ice, Mendenhall Glacier; and young living clear ice from drill core 200 feet below the surface, Taku firn basin, Juneau Ice Field. Without doubt this list is short only because of lack of observation on other glaciers.

Whether this type of structure is rare in rocks and metals or whether it is more common, but has escaped detection, should be determined. Thin-sections will not reveal it; one must resort to repeated grinding and polishing at small intervals, corresponding to small grain size.

The interlocking structure raises the problem of determination of grain size. Isolated rubbings of the Malaspina dead ice suggest an average grain size of only a few inches, which is certainly a great underestimate. Many of the crystals may have maximum linear dimensions of several feet but relatively small cross sections throughout. Linear grain dimensions are not an important descriptive parameter for this type of structure.

From the point of view of glacier mechanics, the structure is of particular interest; it is difficult to visualize shearing along the tortuous grain boundaries as an important factor in grain deformation. In view of the preferred crystal orientations described below, it is tentatively suggested that, in this type of ice, intra-crystalline deformation (lattice translation) is the main flow mechanism, with phase changes according to Riecke's principle (pressure melting and regelation) as important contributors. This question can be solved only by systematic laboratory investigation, never by field work alone.

The length of grain boundary line on each of the 26 illustrated rubbings was measured; the minimum is 60 inches in Nos. *14* and *15* and the maximum is 96 inches in No. *1*. The 320-cubic-inch ice block has a total intercrystalline boundary surface area of 400 square inches. If the grains were cubes, the length of the side would be 2.4 inches for the quoted proportion of boundary surface area to volume. This number might be useful as a descriptive parameter.

The interlocking structure is most probably the result of a fusion of crystals of parallel orientation in a highly preferentially oriented aggregate. In the clear living ice near the foot of the Malaspina, masses of interlocking structure were seen to alternate with masses of simpler structure (more or less equidimensional grains) in an apparently haphazard manner. A block of ice of this simpler structure was also studied by means of equally spaced rubbings. The crystals are somewhat smaller (maximum frequent diameter 6–9 inches against 9–12 inches in the interlocking structure) and roughly equidimensional. Although the degree of interlocking is considerable, it does not wholly dominate the structure. On melting of the intergranular layers, this ice disaggregates into more easily separated single ice crystals. This ice would be an excellent source of large, flawless (relaxed and bubble-free) single crystals for laboratory work on ice mechanics.

The structure of the living bubbly ice described below is of the highly interlocking type, but of smaller grain size ($\frac{1}{2}$–6 inches diameter in section).

GRAIN ORIENTATION

Published opinion on the symmetry of ice largely favors the assignment to a centrosymmetric class, mainly on the basis of theoretical considerations on the positions of the hydrogen ions. This is contrary to morphological evidence, which strongly indicates a polar main axis (class C_{3v} or C_{6v}). This matter should be further investigated; it is conceivable that the polarity of axes may ultimately have to be considered in ice petrofabrics.

In the meantime, determination of grain orientation can be limited to the azimuth and polar angle of the optic axis, although in advanced work it may become desirable also to determine the orientation of secondary axes. The different methods for optical work on ice sections are summarily described below as a guide to prospective workers in this field.

Ice is optically uniaxial and positive, for $\lambda = 550$ (daylight mean), $\epsilon = 1.3120$, $\omega = 1.3106$, and $\epsilon - \omega = 0.0014$.

The dispersion is negligible for our purpose. In sections cut parallel to the optical axis (maximum double refraction), the thickness is 0.40 mm. per order of interference colors; a section 0.8 mm. thick, for example, will show second-order red.

A plane normal to the section and containing the optic axis intersects the plane of the section in a line. The angle between one direction of this line (the direction enclosing the smaller angle with the optic axis above the section) and a reference direction (azimuth zero degrees) is the azimuth. The polar angle is the acute angle between the optic axis and a normal to the section.

PETROGRAPHIC MICROSCOPE WITHOUT UNIVERSAL STAGE

The thin-section is mounted under the microscope on a mechanical stage. The crystal to be examined is brought into the center of the field of view. Four operations follow.

1. *Determination of the thickness of the*

crystal.—The microscope tube must be fitted with a micrometer screw, reading to ±0.01 mm. The top and bottom contacts of the crystal with glass are often recognizable by the presence of small air bubbles. If these are absent, grains of a foreign substance should be smeared on the section before it is put under glass.

The thickness (d) is measured with a high-power objective and is equal to the amount of lowering (d') of the microscope tube on focusing first at the top and then at the bottom of the section, multiplied by the index of refraction of the crystal; hence $d = 1.31d'$.

2. *Determination of the azimuth of the vertical plane containing the optic axis.*—The analyzer is pushed in and the stage rotated to maximum illumination of the crystal. A gypsum plate or quartz wedge is inserted. Subtraction of interference colors indicates that the vertical plane containing the optic axis is parallel to the fast ray of the inserted plate or wedge. The vertical plane is rotated to a left-right position (extinction) and the azimuth read on the microscope stage.

3. *Determination of the general position of the optic axis.*—It is to be determined whether the optic axis is inclined to the left or to the right. This is done conoscopically, the method failing when the polar angle exceeds some 85°. If the conoscopic image is obtained by extracting the eyepiece, it is conformable to the crystal. If the Bertrand lens is used to obtain the image, a 180° rotation is indicated; i.e., if the axis is observed to be inclined toward the left side of the image, it is actually inclined to the right in the crystal.

4. *Determination of the polar angle.*—If the conoscopic image shows the axis to be within the field of view, then the polar angle can be computed from the central distance measured with an ocular micrometer and from the Mallard constant of the optical system.

For polar angles smaller than 15° this method is more accurate than the one described below. A conoscopic image of ice is often fuzzy in sections thinner than 0.3 mm. Weak biaxiality is common.

The other method for obtaining the polar angle consists in measuring the double refraction with a compensator. Owing to the low dispersion of ice, it is unnecessary to resort to monochromatic light. With a Wright eyepiece and graduated quartz wedge it is possible to measure the phase difference (R) to within ±5 mμ; therefore,

$$(\epsilon' - \omega) = \frac{R}{d},$$

$$\sin \rho = \frac{\epsilon}{\epsilon'} \sqrt{\frac{\epsilon'^2 - \omega^2}{\epsilon^2 - \omega^2}},$$

where ρ is the polar angle.

It is convenient to make a graph giving the relation between ρ, R, and d. With R as abcissa and d as ordinate, the values for ρ (for example, at 5° intervals) are represented by straight lines. The graph has been published (Bader, 1938).

The measurements are rather inaccurate for sections thinner than 0.5 mm. At 0.3 mm., the angle of up to 70° may be 3° in error; from 70° to 90° the error increases to from 5° to 7°. The error may reach 10°–15° in sections only 0.15 mm. thick. Thus sections should be made as thick as possible or convenient.

PETROGRAPHIC MICROSCOPE WITH CONVENTIONAL UNIVERSAL STAGE

The petrographic microscrope with universal stage has been successfully used by Robert F. Black of the U.S. Geological Survey. A correction must be made for the large difference between the refractive index of ice and that of the glass hemispheres.

Both this method and the preceding one require the microscopic work to be done in a cold chamber or the microscope stage to be cooled to prevent any melting of the section.

UNIVERSAL STAGE WITHOUT HEMISPHERES

In air.—Plate 1, *B*, shows a stage built for the study of large (9×9-inch) sections of coarse ice. It is a three-axis stage, permitting inclination in only one axis. It was successfully used in a chamber hewn into the dead ice at the fringe of the Malaspina Glacier. Here the temperature was a very few degrees above melting point. Sections 5 mm. thick between glass required 1–2 hours for melting completely.

The main disadvantage of using this stage in air arises from the differences of the refractive indices of air and ice. If α is the measured inclination of the section and ρ the true polar angle of the optic axis, then

$$\sin \rho = \frac{\sin \alpha}{\omega},$$

where $\omega = 1.3106$, the refractive index of the ordinary ray in ice. This greatly limits the measuring range of the three-axis instrument, where the crystallographic *c*-axis must be "stood up." The limit of inclination of the stage is about 70°–80°, which corresponds to a true polar angle of 45°–48°.

The construction of an ice-in-air stage is relatively simple. The top is a horizontal brass ring on which the square frame carrying the upper polaroid ($10\frac{1}{2} \times 10\frac{1}{2}$ inches) can be rotated. The frame with the lower polaroid is suspended from the upper frame by two rods. The stage carrying the ice section has two circles, namely, a brass ring with azimuth graduation on which the section carrier can be rotated and the ring itself, which can be inclined and carries a protractor for reading the angle of inclination. The frame of the instrument is made of light aluminum angle-and-channel beams. The section-carrier is of fiber with two glass plates to sandwich the ice section.[3] The ice section is, of course, wet and tends to slide around between the glass plates. It was effectively stabilized by a dozen pieces (3×3 mm.) of adhesive tape stuck to the lower glass plate.

Before measuring, a sketch of the crystal boundaries of the section is made, unless they have been previously recorded by a rubbing or by other means. A Land polaroid camera would be a great time-saver. In measuring, the following operations are performed:

1. The crossed polaroids are rotated so that their directions of polarization are respectively parallel and normal to the axis of inclination.
2. The section (in horizontal position) is rotated until the crystal to be examined is extinguished.
3. The section is inclined up and down in the horizontal axis. If the crystal lights up, the section is rotated 90°. Inclination will then not cause the crystal to light up.
4. The polaroids are rotated approximately 45°. The crystal normally lights up.
5. The section is inclined to extinction of the crystal.
6. The polaroids are rotated. If the crystal remains dark, then the light is transmitted parallel to its optic axis. Inclination (α) and azimuth are read and recorded.

While measuring, it is important to look vertically through the polaroids. On observing extinction on large crystals, only a relatively small area directly under the eye remains dark. In searching for optic anomalies, the eye is moved around over the whole field of view. If the optic axis lies within some 10° of the plane of the section, it cannot be located,

[3] The two brass rings for the illustrated instrument were kindly contributed by Yale University. They formed part of the original universal stage used by the late Max Demorest.

unless the section is thinner than 2 mm. In this case the successive lowering of interference colors on inclining the section (elevating the optic axis toward the vertical) will show where the axis is. The thicker the section, the longer it will last before melting away, and the sharper the extinction as the optic axis sweeps by on inclining.

Full coverage of all crystal orientations on a single section requires a four-axis stage (two crossed axes of inclination). The crystal axes having more than 45° inclination can then be "laid down" instead of "stood up." Four-axis stages, which are somewhat more difficult to construct than three-axis ones, have recently been used by R. F. Black and R. P. Sharp (private communication). In "laying down" the optic axis, correction must again be made for refraction. If a is the measured inclination of the section and ρ the real polar angle of the optic axis, then

$$\cos \rho = \frac{\sin a}{n}.$$

Whether n is equal to ϵ (extraordinary ray) or ω (ordinary ray) depends on the orientation of the lower polaroid. If this polaroid transmits light parallel to the plane containing the optic axis, then $n = \epsilon$. However, the difference between ϵ and ω is too small to matter in practice.

For information on procedure for "laying down" the axis, see Fairbairn (1949).

In liquid (immersed).—Two important advantages are gained by immersing the universal stage in a liquid. The range of measurement of polar angles of a three-axis stage can be increased by choice of a liquid of suitable refractive index, and the liquid can be cooled to permit work at higher ambient temperatures. This is of particular importance for work in the field, as it saves the dozen or so man-days required to excavate a cool laboratory in névé or ice and the labor of maintenance of the laboratory. Water, surrounded by a jacket of snow or chipped ice, is by far the most convenient liquid to use in the field. If the ambient temperature is too high, causing melting of the section before completion of the measurements, salt can be added to the cooling jacket.

Plate 1, *C*, shows an immersed three-axis stage for circular sections of 2.3 inches maximum diameter. The mechanical and manipulatory complications and the very high cost of a fourth axis are hardly warranted, as polar angles up to 80°–85° can be measured by "standing up" the optical axes of the ice crystals, unless they are very small.

As in the case of stages without hemispheres in air, correction must be made for refraction in the immersed universal stage.

If n is the refractive index of the liquid, then

$$\sin \rho = \frac{n}{\omega} \sin a.$$

If the results are to be plotted on a Schmidt equal-area net, direct plotting of the values of a can be done from an appropriate scale.

If R is the radius of the net, then the distance r from the center of a point corresponding to a polar angle ρ is:

$$r = R\sqrt{2} \sin \frac{\rho}{2} = R\sqrt{1 - \sqrt{1 - \frac{n^2}{\omega^2} \sin^2 a}}.$$

For the standard 20-cm. Schmidt net and for water as immersion liquid ($n = 1.3340$):

$$r = 10\sqrt{1 - \sqrt{1 - 1.036 \sin^2 a}}.$$

Table 1 is computed from this equation.

For laboratory work at temperatures below the melting point of ice, a liquid of higher index of refraction can be chosen to lessen the maximum angle of inclination of the stage. The liquid (butyl alcohol, for example) must, of course, not be a solvent for ice.

For information on the preparation of petrofabric diagrams one can refer to Sander's work, to the more readable excellent summary by Haff (1938), or to the book by Fairbairn (1949). For some reason which I have been unable to unearth, it has become petrofabric practice to project the lower hemisphere, contrary to the crystallographic practice of projecting the upper hemisphere. The projections which follow (Figs. 3 and 4) are of the *upper* hemisphere.

TABLE 1

CONVERSION OF IMMERSED STAGE DATA FOR PLOTTING ON SCHMIDT NET

Measured Angle a (°)	Real Polar Angle ρ (°)	Central Distance r (Cm.)
0	0	0
10	10	1.25
20	20	2.50
30	30	3.73
40	41	4.94
50	51	6.11
60	62	7.27
70	73	8.42
72	76	8.66
74	78	8.91
76	81	9.18
78	85	9.52
79	87	9.80
79½	90	10.00

If an equal-area net cannot be obtained, it is necessary only to draw a 20-cm. diameter base circle graduated at 2° intervals from 0° to 360° for azimuth and a single diameter graduated according to the values in table 1 for polar angles. In order to avoid possible error as to the sense of inclination of the section, it is convenient to graduate the diameter of the circle from 0° to 180° and to have a corresponding scale on the stage to measure inclination. The scale then reads 90° when the section is in a horizontal position.

The special instruments described above are essential only for accurate work; approximate orientation patterns can be determined by simpler means, as follows:

1. The "rotten" ice of the surface of the ablation zone melts in a characteristic manner. More or less cylindrical holes form in the crystals, the cylinder axis being parallel to the basal plane. At an advanced stage of melting, numerous holes join to delineate rough plates, which, on reflection of sunlight, show a striking preferred orientation. The dip and strike of these plates can be measured to obtain azimuth and polar angle. In using this method, care must be taken to avoid biased sampling.

2. In "bubbly" ice most of the bubbles reveal the orientation of the crystal enclosing them. Although the bubbles are commonly round, they are surrounded by water bags, nearly always in the form of a rotational, nonspherical body, commonly a disk. The axis of rotation is invariably parallel to the optic axis of the crystal (Bader, 1950). The difference of index of refraction of water and ice is sufficient to render these bodies easily visible under a magnification of 10–30×.

A flat surface is prepared, and a notebook or, better, a sheet of paper on a wooden board is pinned to the ice to one side of the prepared area. Viewing the water bags with a hand lens or a stereomicroscope, a line is drawn freehand on the paper parallel to the estimated azimuth.

After a little practice it is quite easy

to assign the polar angle with reasonable confidence to one of four intervals of inclination, for example, 90°–80°, 80°–60°, 60°–30°, and 30°–0°. One, two, three, or four arrowheads in the direction of the azimuth can be added to the line on the paper to indicate the assignment.

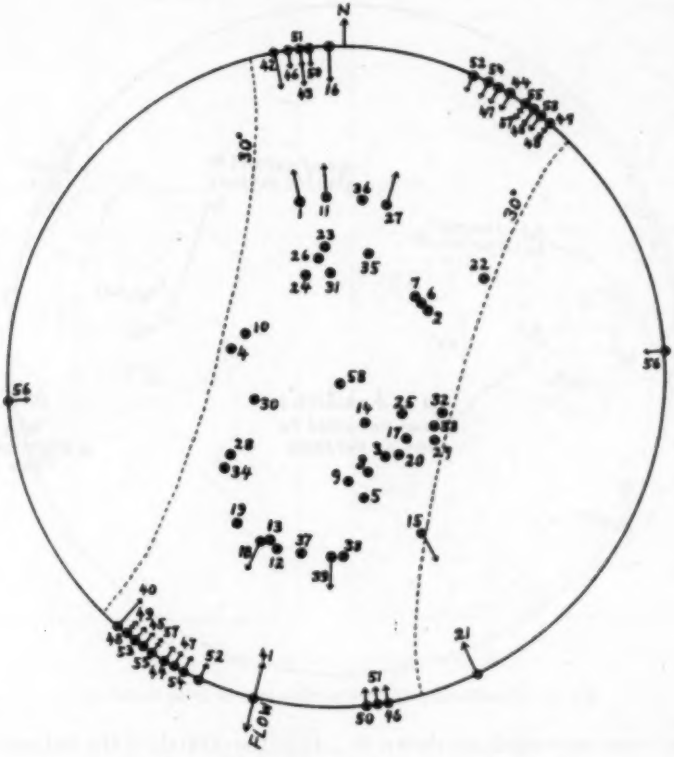


FIG. 3.—Equal-area projection of optic axes of dead ice

3. If the ice is quite clear, then one must have sunshine and a lens (2–4 inches). The sunlight is focused inside the crystal, where it will produce a flat, very thin, Tyndall melt figure similar in shape to a hexagonal snow star. The main axis of the crystal is normal to the plane of the figure.

At present, the Tyndall figure is the only known simple means of determining the relative position of the secondary crystallographic axes—relative because, to my knowledge, the correlation between Tyndall figure and secondary crystallographic axes has not yet been established. (The branches of the figure are either parallel or normal to the axes.)

In its retreat, the Malaspina Glacier has laid down a mighty mass of dead ice along its fringe (pl. 2, *A*), forming an arc some 60 miles long, 3–10 miles wide, and in parts probably as much as 1,000 feet thick. The laboratory at Camp Malaspina was hewn in this dead ice, and the specimen described in detail above was

obtained from the innermost part of the ice chamber. The three sections (fig. 2, rubbings Nos. *26*, *29*, *32*) show crystals Nos. *1–5* (No. *5* is filled in black). Sections of ice from the surroundings of the laboratory were also made as shown in the accompanying tabulation.

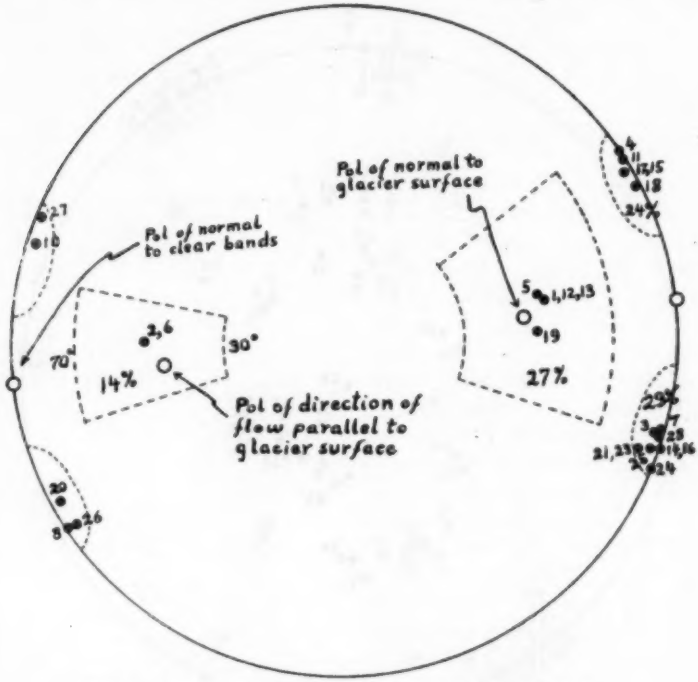


FIG. 4.—Equal-area projection of optic axes of living bubbly ice

Relation to Laboratory	Crystals	No. in Fig. 5
Different parts of lab.	Nos. 6–11	*A* (sketch)
300 yd. north	Nos. 12–21	*B* (rubbing)
500 yd. south	Nos. 22–27	*C* (rubbing)
100 yd. west	Nos. 28–39	*D* (sketch)
300 yd. east	Nos. 40–58	*E* (rubbing)

Bottom to top of the figure corresponds to north in Figure 5, *B–E*, but to N. 30° W. in figure 5, *A*. The 58 crystals, plotted on an equal-area net (fig. 3), constitute a beautiful example of an oriented structure. Arrows pointing outward indicate that the polar angle was greater than 45°. The short arrows pointing inward represent, with one exception, the crystals of the last section listed above. They indicate that the polar angles are all near 90°. An error in the orientation of the block was suspected, but a check showed that it was correct. Thus there appear to be significant local variations in the orientation patterns of the ice. All taken together, the main axes of the crystals lie outside a cone with apex angle of approximately 120°. Within this cone there is a paucity of crystal axes near the vertical and near the horizontal parallel to the direction of flow. The dead ice body is here at least

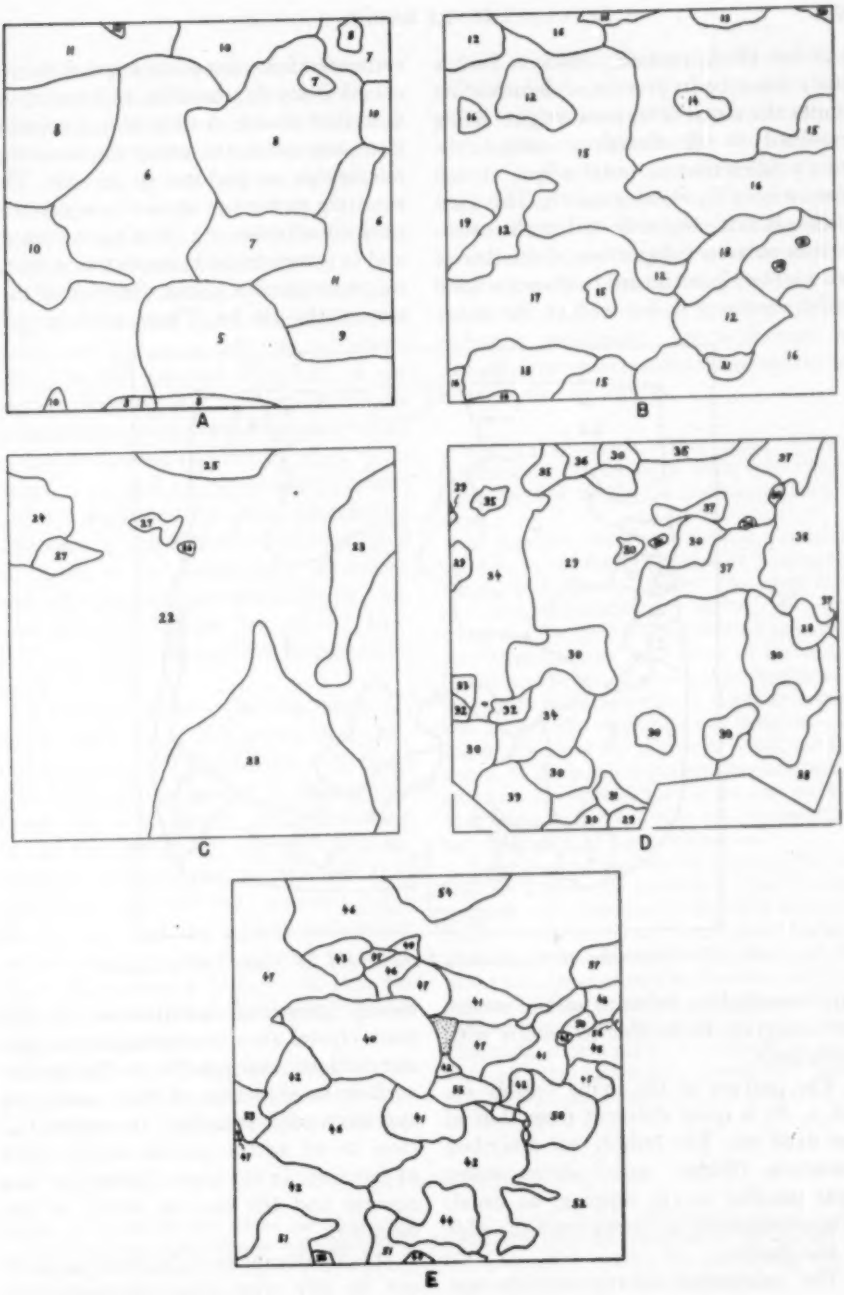


FIG. 5.—Thin-sections of dead ice. Frames 8 inches square

150 feet thick, probably thicker. Such a body must be in process of deformation under the stress of its own weight. Being confined in all directions except the south (direction of outer edge), it can flatten only by flowing south. This condition is evidently reflected in the orientation pattern (proportion of number of axes inclined southward to those inclined northward is 3:2), but we lack the necessary correlations between stress, strain, and structure to be able to read it with confidence.

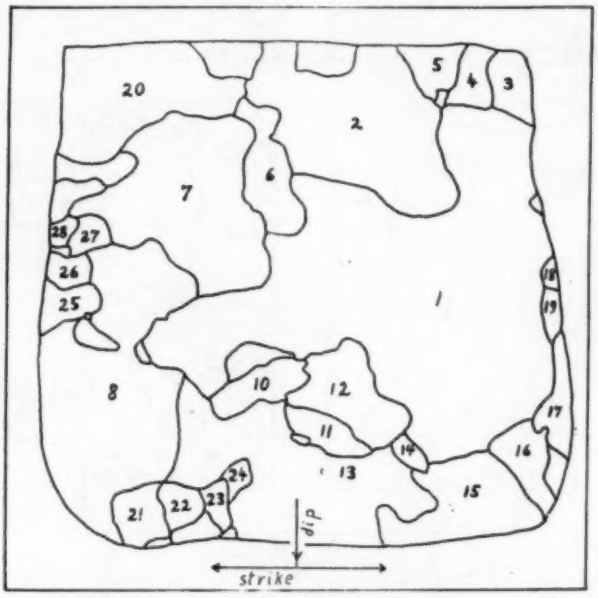


FIG. 6.—Thin-section of clear band in living bubbly ice. Frame 8 inches square

The pattern of the living bubbly ice (pl. 2, *B*) is quite different from that of the dead ice. The bubbly ice described elsewhere (Bader, 1950) shows many clear parallel bands, dipping, as usual, at approximately 45° away from the edge of the glacier.

The orientation of the crystals was estimated from the orientation of the enclosed disks (air bubbles and water) as described above. A total of 235 orientations was estimated under the binocular microscope on perhaps 50 crystals. The resulting pattern is shown in equal-area projection in figure 4. It is most striking and is symmetrical in respect to a vertical plane parallel to the direction of motion of the glacier. There are four very clearly preferred orientations of the main crystal axes; one normal to the glacier surface, one parallel to the glacier surface in direction of flow, and two symmetrically located at angles of close to 20° to the normal on the plane of banding, in the plane formed by this normal and the line of strike of the banding.

It is significant that the clear bands do not in any way form distinguishable

structural units, except that they are clear, nearly bubble-free, flat zones in the milky mass of bubble-rich ice. The boundary between clear and bubbly ice is usually quite sharp but does not, in general, follow grain boundaries; nor do the crystals in or enclosing the clear bands form a different orientation pattern.

This is demonstrated by a thin-section (fig. 6) of a clear band. The numbered points on the projection (fig. 4) correspond to the numbered crystals of the section. Here again it is impossible to determine the number of individual crystals in the illustrated section. Crystals numbered *1*, *12*, and *13* are optically parallel but are separated by grain boundaries. The same is true of those numbered *2* and *6*. Owing to the interlocking structure, optically parallel elements without common boundary (*14*, *16*—*15*, *17*—*21*, *23*—*5*, *1*—*3*, *7*) may or may not be parts of a single crystal.

It appears unlikely that the bands are active shear zones, if it is true that crystals tend to orient themselves with basal translation planes parallel to shear planes (even this is doubtful). The temptation to call the clear bands relics of old shear planes is counteracted by the fact that they have their "correct" position in the glacier and that the crystal orientation pattern is very clearly related to them. Here again there is little doubt that the solution to the puzzle is implicit in the orientation pattern.

It would be premature to discuss stress-strain–orientation-pattern relationships before a great deal of further data becomes available. Reasoning by analogy with metals and rocks is possible but rather unconvincing. Wet ice is essentially a two-phase system, never far removed from the equilibrium state because of a high power of relaxation. The deformation mechanism of slippage along grain boundaries, crystal lattice translation and regelation tends to flatten stress gradients more or less toward hydrodynamic conditions. Ice is perhaps as unique among crystalline aggregates as water is among liquids.

ACKNOWLEDGMENTS.—This paper in part summarizes the results of work done during eight weeks in the field in Alaska in 1949 and 1950.

In 1949, five weeks were spent at a camp located on a sand flat on the dead ice fringing the Malaspina Piedmont Glacier. The camp location was approximately 140°25′ W. and 59°43′ N. The camp was established as part of the "Snow Cornice" expedition of the Arctic Institute of North America and partly financed by a grant from the Office of Naval Research, U.S. Navy. The expedition was under the expert leadership of Mr. Walter A. Wood, and Dr. Robert P. Sharp ably organized the glaciological program. My special thanks are due to Mr. Alan Bruce-Robertson, who was my assistant and companion at Camp Malaspina.

In 1950, three weeks were spent on the Juneau Ice Field, some 30 miles north of Juneau, at Camp 10B of the Juneau Ice Field Research Project of the American Geographical Society. This expedition was led by Mr. Maynard M. Miller and also supported by the Office of Naval Research. The project supplied funds for the construction of the immersed universal stage illustrated in this paper. The results of work done with this instrument on drill cores from 150 to 300 feet in depth will be described in another paper.

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TEMPERATURE OBSERVATIONS IN BERGSCHRUNDS AND THEIR RELATIONSHIP TO CIRQUE EROSION[1]

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ABSTRACT

Willard Johnson's bergschrund hypothesis assumes diurnal temperature changes at depth. Direct evidence from continuous temperature records in bergschrunds calls for a modification of this hypothesis. Evidence from northeastern Greenland and Norway shows that air temperatures in an open bergschrund during the summer vary between 30° and 32° F. In Norway diurnal changes of temperature were absent in spite of substantial changes in air temperature outside. New continuous recording instruments have been designed to measure temperatures and the number of times the temperature crosses the freezing point. Further direct evidence is necessary before adopting a thaw-freeze hypothesis.

Willard D. Johnson's (1904) justifiably renowned descent of a bergschrund and his careful account of what he saw opened a new era in the study of cirque erosion. It is scarcely an exaggeration to say that it heralded a change of emphasis from description of the characteristic features of glaciated mountain scenery to the investigation of the processes responsible for these features. The inferences he somewhat cautiously drew from his observations were taken up by his followers with an enthusiasm that embarrassed him. According to Isaiah Bowman (1920, p. 296), Johnson, before he died, had fully intended launching an attack on the so-called "bergschrund hypothesis."

Johnson's observations are important primarily because they pointed to thaw-freeze processes operating deep down in a bergschrund, but he also revealed the true nature of a bergschrund. The bergschrund he descended continued down until it reached rock on the up-glacier side. The *randkluft*, on the other hand, which often forms late in the melt season, is a crack wholly between the rock wall of the cirque head and the névé. The distinction, though useful, is somewhat academic, as what appear to be bergschrunds early in the melt season frequently turn into *randklufts* later on, especially at the end of seasons in which ablation has been excessive. Whether or not one should refer to a crevasse near the head wall as a bergschrund, as Bowman did, even though it does not reach bedrock, is as yet unsettled. It is probably wisest to keep the term "bergschrund" fairly wide, as, in fact, the crevasse or crevasse series to which the name was originally applied varies much from place to place and from time to time. It resembles a living organism with a life-cycle concerning which we are extremely ignorant.

The great advantage of the bergschrund hypothesis was that it provided a simple mechanism of attack by a glacier upon the rock wall at its head. The day-to-day changes of temperature across the freezing point which are so effective in shattering rocks lying high up on the mountain sides above the blanketing snow or ice have, nevertheless, still to be demonstrated deep down in such crevasses. Lewis (1940), having been largely initiated into the study of snow and ice in the exceedingly moist climate of southeastern Iceland, considered that meltwater must aid substantially the process envisaged by John-

[1] Manuscript received April 23, 1951.

son and his followers. He repeatedly observed meltwater descending into glaciers and névé fields, especially where any rock rose above the surface of the ice. This modification helped to overcome the rather serious objection to Johnson's theory that warm air cannot displace the heavier cold air occupying a crevasse, except under distinctly abnormal conditions. Equally, direct or indirect radiation from the sun cannot transfer much heat a hundred or more feet down a narrow crevasse in which air circulation and radiation alike tend to be seriously hindered by snow bridges. Furthermore, severe erosion of the rock wall was examined by Lewis in Switzerland *above* a bergschrund, where the tunnel from the Jungfraujoch railway station within the mountain emerges on to the Jungfraufirn. An examination of British cirques also suggested that erosion by glacial plucking and shattering—for which thaw-freeze seemed to offer the best explanation—occurred right to the base of 1,000-foot head walls. Thus it was argued that an erosive process similar to, if not identical with, thaw-freeze had once been active over zones of cirque head walls many hundreds of feet in vertical extent.

However, the meltwater hypothesis, like its parent-hypothesis, was based on somewhat casual observation and inference, and the search for definite evidence is overdue. Johnson, at least, was quite honest in admitting that he "was at the disadvantage for close observation of having to clamber over these [blocks] with a candle, in a dripping rain of meltwater." Battle therefore set himself the task of searching for more direct evidence of what really happened in bergschrunds by emulating Johnson's feat of penetrating deeply into a bergschrund, but armed with instruments for measuring temperature, preferably over long periods. This seems to be a new line of attack, perhaps owing to its arduous nature. Only a very few such attempts have been made previously, as far as we know. Von Klebelsberg (1920, 1948) examined bergschrunds encountered during tunneling operations on the Austrian-Italian frontier during the 1914–1918 war. He noted the absence of melting such as would be expected if warmth penetrated the bergschrunds according to the then widely held views. McCabe (1939, p. 463) recorded summer air temperatures in fairly shallow bergschrunds in Spitsbergen; he found them to be usually 0° C., even when the air temperatures outside were far higher. D. H. Chapman sent a letter in 1948, stating that his student, Robert Lange, took temperatures 50 feet down a bergschrund in Teton Glacier, Wyoming, with a thermograph, for most of a period of 3 weeks in the summer. These observations were, we understand, of a somewhat preliminary nature and have not yet been continued.

Battle (1949), as leader and glaciologist of the Leeds University Greenland Expedition in 1948, was able to examine

PLATE 1

A, Grif Gletscher, the corrie glacier above the Pasterze Glacier, northeastern Greenland, August, 1948. The bergschrund is just visible as a thin line below the head of the glacier. Photograph by W. R. B. Battle.

B, The interior of a 30-foot-deep bergschrund in Hellstiggutind, Jotunheim, August, 1950. The transition from névé to ice is plainly visible. The dirt band at the base of the névé marks this transition. The irregular surface below consists of ice masking the rock wall. Photograph by W. R. B. Battle.

C, Large blocks of gabbro broken loose from the head wall of Juvass Kjedlen, August, 1947. In a normal year this whole area is covered by the névé of the cirque glacier. Photograph by W. V. Lewis.

B

A

C

Cirque features in Greenland and Norway

the bergschrund of a corrie glacier[2] on the mainland of northeastern Greenland near Clavering Island (pl. 1, *A*). During the day the bergschrund, which was only barely visible on the surface as a narrow crack, had meltwater trickling down the névé and ice on the up-glacier side. This, however, seemed to be localized at one or two small streams. At the bottom, about 100 feet down, the bergschrund opened out into a huge cavern with a roof of snow blocks. Even at night, when air temperatures in the open were as low as 23° F., it was comparatively warm beneath, where some of the icicles continued dripping. The rock wall appeared in only one or two places as a smooth black slab, mostly covered with a coating of transparent ice, 2 inches or more thick. It was quite impossible to tell whether the floor reached bedrock, as it was covered by a confusion of large blocks of ice fallen from the roof nearly 100 feet above.

The almost uniform cover of ice which remained day and night during the 10 days of observation at the height of summer was surprising and suggests that relatively little damage was being done to the rock face by thaw-freeze or any other process. There were no signs of diurnal changes of temperature. Conditions of melting and of the unfrozen state of one's outer clothing remained unchanged day and night, as far as one could judge without thermometers. If these conditions—seemingly harmless to hard rock—are in any way typical of summertime, it is interesting to speculate as to when the cirque was formed. Lower temperatures, such as occur in winter, would produce little or no meltwater; only relatively hot or wet spells in summer would presumably produce sufficient water to melt through to bedrock. The summer of 1948 was a relatively warm one, and when the party left on August 28 the winter freeze-up had begun. One was left with the strong impression that erosion was not active under present climatic conditions. It would have been even less so half a century earlier, when Koldeway (1873) records that the ice covered everything except one or two nunataks. Little is known about the previous extent of the ice, but Ahlmann (1948, p. 166) and Brooks (1949, p. 357) suggest that it was much less extensive during the period of the Viking discoveries from 900 to 1400. Cirque erosion may have been more active in Greenland then.

These indeterminate but thought-provoking observations in Greenland stimulated the co-author to continue his work in the more accessible Jotunheim, where other glaciological field work was being undertaken by the Cambridge group. The instruments available for this work in the summer of 1950 were a mercury-in-steel Negretti and Zambra thermograph with a 10-inch bulb, very kindly loaned by Professor Gordon Manley, of Bedford College, London, and six small radioactive temperature recorders designed and loaned by Dr. Eilif Dahl, of the botany department of the University of Oslo. The thermograph was a standard instrument driven by an 8-day clock. Its accuracy to about 0°.3 F. satisfied our requirements, but it was distinctly cumbersome, weighing 30 pounds. Also, it needed visiting each week—a wise precaution for such work, as any minor mishaps to the instruments can then be promptly remedied.

The Dahl (1949) recorders are extremely ingenious and stoutly constructed devices, measuring only $6 \times 3 \times 1\frac{1}{2}$

[2] The Danish State Department Committee on Greenland Names has just authorized the suggested name for this glacier, "Gryphon Gletscher," in the Danish form, "Grif Gletscher."

inches and weighing only a pound or so. A bimetallic spring moves a radioactive source, which emits a sheet of alpha particles on to a small photographic plate. The length of time for which the temperature remains at any particular value is estimated from the intensity of the record on the photographic plate. These recorders were designed for botanical work, such as measuring the duration of the growing season in the distant coniferous forests of northern Norway. They possess the great advantage of being able to be left out for a period of up to 3 months. This made them most convenient for use in bergschrunds. One was placed alongside the thermograph for standardization, and the others were placed in distant bergschrunds, such as that in one of the left-hand tributaries to Heillstuggubreein (pl. 1, *B*). The chief limitation for our purpose is that these recorders do not indicate the number of times the temperature crosses the freezing point but only give the approximate length of time at various temperatures.

The thermograph record was made at the bottom of a 30-foot bergschrund at the head of Tverrabreein, Midt-Jotunheim (6,875 feet.).[3] The bergschrund was narrowly open to the air, the gap widening to 3 feet at the surface by the middle of August. The bulb was fixed within 3 inches of the rock wall, which had a slight covering of glazed frost on it and was remarkably dry. The chart was changed three times, but, owing to other demands on the personnel for the varied glaciological program, the changing of the chart was unavoidably delayed the first week and the pen ran dry. With this 2-day break, the record ran for 26 days. In figure 1 the maximum and minimum temperatures and wind speeds (Beaufort scale) at the meteorological station on Fannaraaken, less than 14 miles to the west and only 100 feet lower, are added. Both temperature scales start at 29° F., but the Fannaraaken record, on the right-hand side, is reduced four times compared with the bergschrund temperatures, given on the left of the figure. Comparisons of temperature on the meteorological station at the Jungfraujoch, Switzerland, which, like the Fannaraaken station, is situated on a peak, with those taken on the glacier below by Seligman's party in 1938, were found to be somewhat erratic. Therefore, too much confidence must not be placed in the Fanaraaken records as an indication of air temperatures on the surface above the Tverrabreein bergschrund. Also the standard screen which records temperatures four feet above the ground is not well suited to measuring the temperature of what may be no more than a shallow downdraught of air spilling over the rock wall of the cirque. This cirque culminates in the peak of Bukkehöe, 800 feet higher than Fannaraaken. However, lacking meteorological records taken at the bergschrund, we are fortunate in having them in the same mountain mass and at a similar altitude.

The truncation of the curves at the freezing point is a striking and significant feature of the graph of bergschrund temperatures. This occurred in spite of the fact that the outside air temperature rose to 45° F. at Fannaraaken. It seems that the peak temperature in the bergschrund is determined by that of the surrounding ice, which, by late July, would be at 32° F. This fact is extremely important, and it is supported by the preliminary results of the Dahl recorders placed in other bergschrunds and which Dr. Dahl has kindly just forwarded to us. This fact alone must go a long way to-

[3] See 1:50,000 map, Midt-Jotunheimen, 1935 Norges Geografiske, Opmåling.

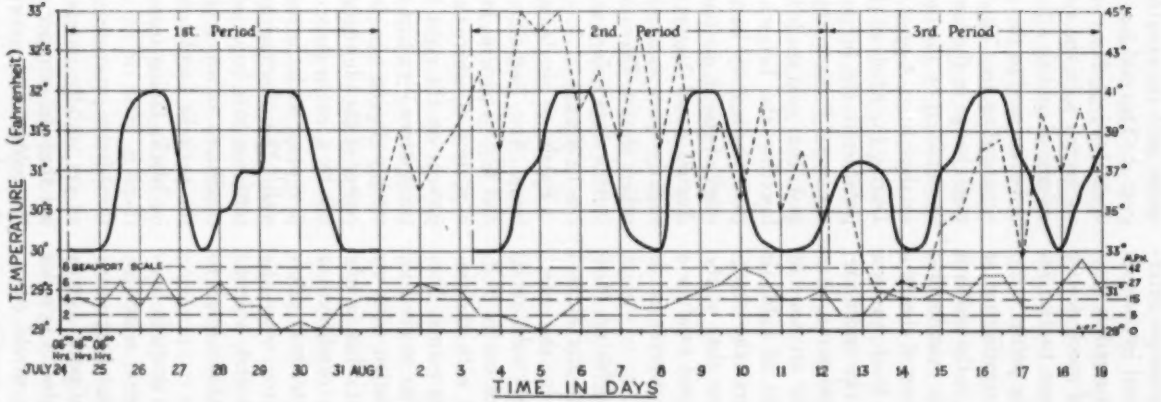
1st. Period
2nd. Period
3rd. Period
TEMPERATURE (Fahrenheit)
BEAUFORT SCALE
JULY 24 25 26 27 28 29 30 31 AUG 1 2 3 4 5 6 7 8 9 10 11 12 13 14 15 16 17 18 19
TIME IN DAYS

FIG. 1.—The continuous line with scale on the left-hand side is the record of air temperatures at the bottom of the 30-foot bergschrund. The broken line indicates air temperatures at these times at Fannaraaken, the nearest meteorological station; the scale is on the right-hand side. The wind speeds at Fannaraaken are shown by the dotted line at the foot of the graph.

ward disposing of the oversimple interpretation of the bergschrund hypothesis of Willard Johnson, which requires temperatures at the bottom of much deeper bergschrunds than this one to rise repeatedly above the freezing point.

A second feature, the significance of which is not yet apparent, is the uniform lower limit of 30° F. Close examination of the instrument trace showed that the uniformity of this lower limit was not quite so marked as that of the upper one. There were two falls of temperature to nearly 29°.5 F., lasting for an hour or more. Why the indraught of cold air which presumably lowers the bergschrund temperature below that of the enclosing walls of ice did not produce a more variable fall of temperature is puzzling.

Perhaps the most striking feature of all is the complete absence of any diurnal change of temperature in the bergschrund, in spite of the very substantial changes at Fannaraaken of nearly 8° F. This, again, is consistent with the impressions which Battle has gained from all the bergschrunds he has worked in. This lack of direct correlation with air temperatures suggests that other factors may operate, such as calm conditions or falls of pressure. The former may allow a stream of cold air to pour down the back wall into the bergschrund, whereas the latter may encourage this cold air to be sucked in.

The correlation between drops in temperature and the prevalence of relatively still air conditions is at least suggestive. Fannaraaken is an exposed peak. Let us assume, therefore, that a wind of 15 miles per hour at Fannaraaken would correspond with relatively calm conditions in the more sheltered vicinity of the bergschrund. Then it would seem that each fall in temperature occurred during calm spells, with the notable exception of August 10. This still leaves unexplained the undoubted 3½-day cycle, which may be connected with the time it takes for the volume of cold air filling the bergschrund to be warmed to the temperature of the surrounding ice. This cyclical temperature change in the bergschrund seems to be modified by distinct cold spells, such as that from August 11 to August 15. During the whole of this period the temperature in the bergschrund rose barely above 31° F. Even this short record suggests that cold spells have more effect in lowering the bergschrund temperature than warm spells have in raising it. The marked rise in air temperature between August 4 and August 6 produced only a slightly more marked "peak" than the others shown. There is patently need for further data before any firm conclusions can be drawn.

Battle, in his descents, nearly always found the bergschrund walls frozen and dry. A general impression is shown in plate 1, *B*, where the rock wall may be present on the right beneath a layer of transparent ice a foot or more thick. The steeply sloping névé layers shown in the center of the photograph are also underlain by a foot or more of hard ice actually in contact with the ice covering the rock wall. The thermograph record further confirms this impression of dry, frozen conditions, in which all is inert except for the periodic bombardment of névé and ice blocks falling from the roof. In fact, there is little or no direct evidence of thaw-freeze processes operating. Lewis is, as yet, mildly unrepentant and points out that the summer of 1950 was one of the coldest and snowiest for years and that therefore melting must have been far less than usual at that season and the bergschrunds very much less open. In fact, the meltwater variant of the berg-

schrund hypothesis requires that temperatures in bergschrunds should be below the freezing point. Battle's experiences in silent bergschrunds, where no dripping or trickling of water could be heard, are contrary to Lewis' experiences in shallower bergschrunds and on the glacier surfaces in névé regions. The whole may well help to build up a truer picture of the variations that can occur from locality to locality and from season to season. It is abundantly clear, however, that no preconceived views must be held overtenaciously in the face of new and important data.

Battle reports that his Easter, 1951, observations in the bergschrunds of the Jungfraufirn, Switzerland, confirm his impression of the low tempratures that generally prevail in Jotunheim bergschrunds. On this occasion the thermograph was placed at the bottom of a 100-foot bergschrund which was almost totally enclosed at the top. In this the temperature wavered between 26° and 28° F., and again all seemed frozen and still. We await with increased interest the complementary observations which he proposes to make at the same localities in the autumn of this year. He has buried several recorders, designed by J. W. Glen, of the Cavendish Laboratory, Cambridge, and himself, which indicate on post-office counters the number of times the temperature crosses the freezing point or any selected point a few degrees lower, according to the freezing mixture used.

A trial test is also being made with a new thermograph designed by J. E. Jackson, a lecturer in the department of geodesy and geophysics, Cambridge, and Lewis. It is based on Dahl's design, in that temperature changes affect a bimetallic spring which carries a needle point. This scratches a thin mark on a circular glass plate covered with lamp black. The plate is turned slightly each time the temperature rises and is held stationary each time it falls. So the trace should be recognizable even if the instrument is left unattended for a year or more. Unfortunately, we shall not know the date of any temperature changes recorded. An improvement which will give a finer record, with no friction on the moving parts, and a rough time scale is being developed with the kind assistance of Dr. D. H. Wilkinson, of the Cavendish Laboratory. This will replace the needle by a fine jet of alpha particles from a radioactive source attached to the free end of the bimetallic spring, and the lamp-black by a photographic plate. We hope to arrange for the alpha particles to emerge at an average frequency of one per hour. This will give an approximate time scale, either by counting the number of marks on the photographic plate or, at worst, by estimating the intensity of any blurs produced while the plate was stationary.

These comparatively low temperatures recorded in the bergschrunds form an instructive comparison with those which the Swiss engineers (Haefeli, 1951, and personal communication) experienced at the rock-wall end of the tunnel through the foot of the Mount Collon ice-fall of the Lower Arolla Glacier. There, far below the névé field, the winter temperatures were always near the freezing point, and the workmen were loath to emerge into the far colder air outside the tunnel. In such circumstances it is difficult to imagine how the water, pouring down the rock step beneath the glacier, could ever freeze. Battle's experience in bergschrunds with temperatures slightly below the freezing point also leads him to doubt whether freezing readily occurs. So the best evidence of thaw-freeze is still

the shattered nature of rocks newly exposed.

Plate 1, *C*, shows the destruction caused by thaw-freeze in the strong gabbro in the head wall of Juvashytte cirque, Mid-Jotunheim. Great blocks of rock, many weighing tons, were wedged loose —and pushed slightly away—from the rock wall. The site is normally covered by névé, but the photograph was taken in the summer of 1947, when phenomenal ablation followed a winter in which little snow had fallen. Professor Faegri, of Bergen, reported that the snouts of many Norwegian glaciers receded 30 meters a month during that summer. Thus we were given a glimpse of bedrock which had probably not been exposed for a thousand years or more.

Figure 2 records the mean monthly maximum and minimum temperatures at Fannaraaken (1933–1946) and at Ben Nevis (1884–1903). The minimum at Fannaraaken is below the freezing point except in July, so this is almost always low enough to allow freezing to occur. On the other hand, the maximum is above the freezing point only in June, July, and August and is very slightly below in May and September. Does this mean that at Fannaraaken it is possible ordinarily for thaw-freeze to operate only in the three summer months of the year? Priestley (1923, p. 33) records that thawing occurred in the Antarctic when the air temperature was below the freezing point, provided that there was much black surface rock or debris about. This may mean that limited thaw-freeze can occur in May and September and occasionally in colder months in the Jotunheim.

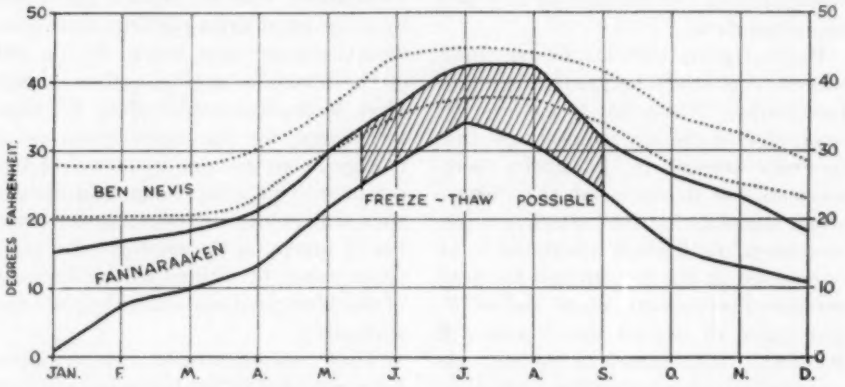

FIG. 2.—Mean monthly maximum and minimum temperatures at Fannaraaken, 6,763 feet (1933–1946), and Ben Nevis, 4,406 feet (1884–1903), showing approximate period when freeze-thaw could take place.

The Ben Nevis results form an interesting comparison. Ahlmann (1948) and Manley (1949, p. 190) have both estimated that this mountain is only a few hundred feet too low to be glaciated today. The Ben Nevis temperatures suggest that thaw-freeze might operate in April and May and again in September and October.

These two examples give no more than a rough indication of when thaw-freeze may operate and when it cannot. In conclusion, it should be stressed that the range of air temperatures, which must influence in a subdued form the changes of temperature in bergschrunds, varies

much from place to place today. In summer the diurnal range is small in Greenland, moderate in the Jotunheim, and considerable in the Alps at the respective altitudes of the névé fields.[4] It would thus appear that the opportunity for freeze-thaw is greater in lower latitudes than nearer the poles. Furthermore, it is clearly greater in shallow, open bergschrunds than in deep, confined ones. An increasing number of fluctuations of climate is being revealed as our knowledge of the Pleistocene expands. We must therefore allow for considerable changes occurring over the centuries in the conditions associated with any given bergschrund locality. The absence of frost-shattering today in a particular bergschrund does not necessarily prove that it has not occurred previously. It certainly induces caution in advocating thaw-freeze, but our real lesson should be to continue with our researches.

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[4] The following absolute maximum and minimum temperatures were recorded on the Jungfraufirn in 1938 by Seligman's party (unpublished monograph by E. A. Ferguson):

Month	Max. (° F.)	Min. (° F.)
May	60.8	7.7
June	72.0	14.0
July	66.2	14.0
August	63.5	19.4

ROTATIONAL MOVEMENT IN CIRQUE AND VALLEY GLACIERS[1]

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ABSTRACT

The concept of rotational slip in glaciers, previously introduced by Lewis, must now be considered in the light of recent theoretical work by Orowan and Nye, based on the assumption that ice is an ideally plastic material and that rate of stress is a function of the shear stress. This paper is chiefly concerned with the formation of shear surfaces in "passive" flow. Possible inferences of Nye's work are discussed. If substantial differential movements develop along the surfaces of maximum shear stress of the passive state, we may have found an important part of the explanation of the variation of intensity of glacial erosion.

Previous work relating to rotational movement is surveyed; and important pioneer work by Gibson and Dyson is discussed.

Results of systematic studies of banding of various cirque glaciers in the Jotunheimen area of Norway are presented. The remarkable systematic behavior of the dips of the bands, which increase steadily away from the head wall and decrease steadily again toward the terminal moraine, is most easily attributable, at present, to some form of rotational movement.

Surface features of the ice below the Mount Collon icefall are discussed in relation to subglacial conditions exhibited in a tunnel dug through to the rock wall. Possible implications of a set of data concerning ice movement and distribution of banding are discussed. We may have here the first evidence of rotational movement beneath an icefall.

INTRODUCTION

Recently Lewis (1949*b*) introduced, perhaps somewhat uncritically, the hypothesis of rotational slipping in glacier flow. It was assumed that (*a*) in a cirque basin, (*b*) at the foot of an icefall, and (*c*) at the snout of a glacier, the body of ice, under favorable circumstances, has a rotational movement. The axis of rotation is horizontal, transverse to the valley direction, and is some distance above the surface of the glacier. In some cases, when the rock bed is slightly bowl-shaped, the glacier was envisaged as having rotated on this bed. In other cases differential movement was assumed to have occurred along curved shear surfaces either within the body of the ice or along the bed of the glacier. It was assumed that such rotation, when it occurred, formed only a part—and in most cases a small part—of the total movement of a glacier.

The evidence for rotational movement was indirect. It consisted of (*a*) the angle at which slabs of rock emerge at the surfaces of glaciers and (*b*) the angle of dip of the ice bands of which glaciers are composed. Following a suggestion by Professor Orowan, we use the terms "bedded bands" for ice bands which originated in the firn region and "flow bands" for those bands formed lower

[1] Manuscript received May 24, 1951.

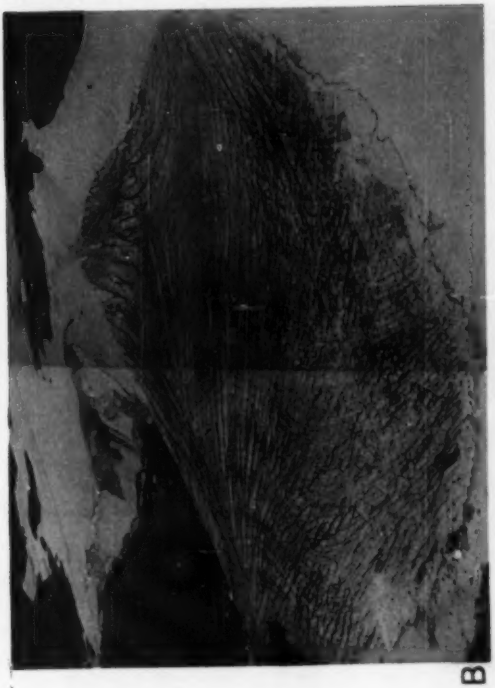
PLATE 1

A, Foreshortened view of Skauthöebreein, Mid-Jotunheim, August, 1949, showing the pattern of ice bands, together with rock debris, outcropping on the surface of the glacier.

B, Juvassbreein, Mid-Jotunheim, August, 1949, showing the pattern of the ice bands and crevasse systems.

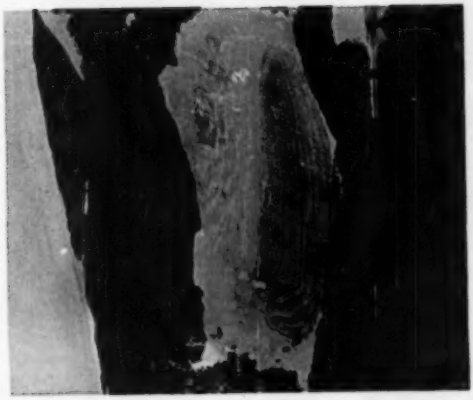
C, The 80–100-foot ice cliff of Juvassbreein, July, 1948, showing the arrangement of curved ice bands as seen in section roughly parallel to the direction of movement of the glacier.

D, Juvassbreein ice cliff, July, 1948, to the right of the section shown in *C*. Note the straightness of the bands and the general decrease of dip as the terminal moraine is approached.



B

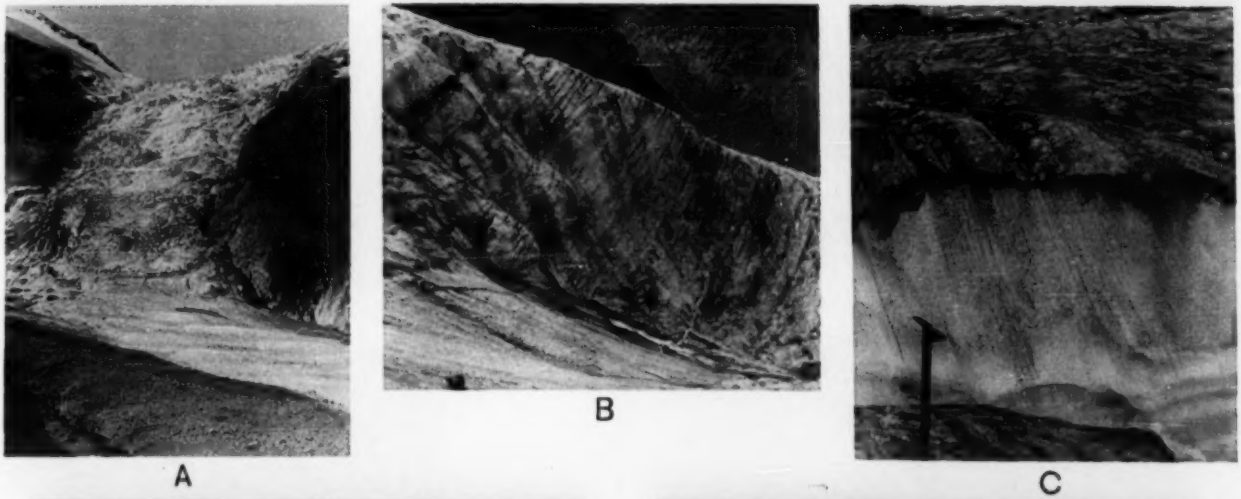
D



A

C

Ice bands of Norwegian glaciers



Features of Mt. Collon ice fall and Lower Arolla Glacier

down the glacier by shearing or by other processes associated with the movement of the glacier. The general nature of these bands is well shown in plan in plate 1, *A* and *B*, and in section in the 80–100-foot ice cliff of Juvassbreein (pl. 1, *C* and *D*.

THEORETICAL CONSIDERATIONS

The whole matter of rotational movements in glaciers has now been put on a sound theoretical basis by Dr. J. F. Nye (1951) of the Cavendish Laboratory, Cambridge, who started this work with a group under the direction of Dr. Orowan. Mr. J. W. Glen, also a member of this group, is carrying out laboratory experiments on the mechanical properties of ice.

Orowan has pointed out that it is wrong to treat ice, as many do, as if it had a constant coefficient of viscosity; for, like all polycrystalline materials near their melting point, it shows the phenomenon of viscous creep with non-Newtonian characteristics. Orowan made the assumption, for mathematical convenience, that ice is an ideally plastic material, i.e., that it remains rigid until a critical stress is reached, which is then sufficient to produce any rate of plastic strain. Orowan showed that, on this assumption, a block of ice would slip down a slope without deformation within the mass, provided, of course, that the thickness of the block and the gradient were sufficient. If a more general assumption is made, that the rate of stress is a function of the shear stress, then the upper layers of the block will move faster than the lower layers. This gives what is probably the normal type of glacier movement, which has been so strikingly demonstrated by Perutz' (1950) velocity measurements in the bore hole through the Jungfraufirn. He placed a steel pipe in a vertical bore hole which extended the full depth of the névé field: he took inclinometer measurements of the pipe and also recorded the movement of the pipe at the surface over a period of two years. This is the first direct measurement of the velocity of a glacier throughout its depth.

Nye (1951) made the same two assumptions but obtained two additional solutions. He termed them "active" and "passive" flow, by analogy with the active and passive Rankine states in soil mechanics (Terzaghi, 1947). The active state involves a downward movement of the surface ice, such as occurs in the accumulation area and especially where a glacier is extended, as near the top of an icefall. The passive state, on the other hand, involves an upward movement of the surface ice, such as occurs in the ablation area and especially where compression occurs near the foot of an icefall below which a long, gently sloping tongue continues.

Figure 1 sums up Nye's theoretical findings concerning the location of surfaces of maximum shear stress within an idealized glacier. These trajectories of maximum shear stress, or families of slip

PLATE 2

A, The Mount Collon icefall and the Lower Arolla Glacier, Switzerland.

B, Ice and dirt bands immediately below the Mount Collon icefall.

C, The steeply dipping ice bands of the Lower Arolla Glacier as seen in the side of a stream channel.

D, Transverse bulge at the foot of the Mount Collon icefall, which has reversed the drainage on the glacier surface.

E, A larger stream maintained its course through the rising transverse bulge shown farther to the left in *D*.

lines, fall into two sets, shown by full and broken lines, respectively. Each set intersects the surface of the glacier at 45° and follows sweeping cycloidal curves down into the glacier, the one set, marked by the firm lines, meeting the bed at a tangent and the second set, marked by dotted lines, doing so at right angles. Substantial differential movement, or shearing, can clearly occur only along the set which meets the bed of the glacier tangentially. Movements along the surfaces, represented by broken lines, can be only in the nature of minor adjustments. However, the theory assumes that shearing movements are small by comparison with the general downhill movement of the glacier.

In applying this theory to glaciers, the slope must not change much in distances equal to the thickness of the ice. In the cirque basins at the head of the glacier and near the head of each icefall, the effective shear surfaces of the active Rankine state dip down-glacier. They may be represented in glaciers by the block-faulting frequently shown by the uppermost seracs at the head of icefalls. In this paper we are more immediately concerned with the shear surfaces in passive flow which occur down-glacier from the foot of steep slopes and icefalls. The effective shear surfaces, in this case, leave the rock bed tangentially and curve upward to intersect the surface at 45°. If shearing and overthrusting occur along these surfaces, then the overlying ice masses concerned exhibit a rotational movement.

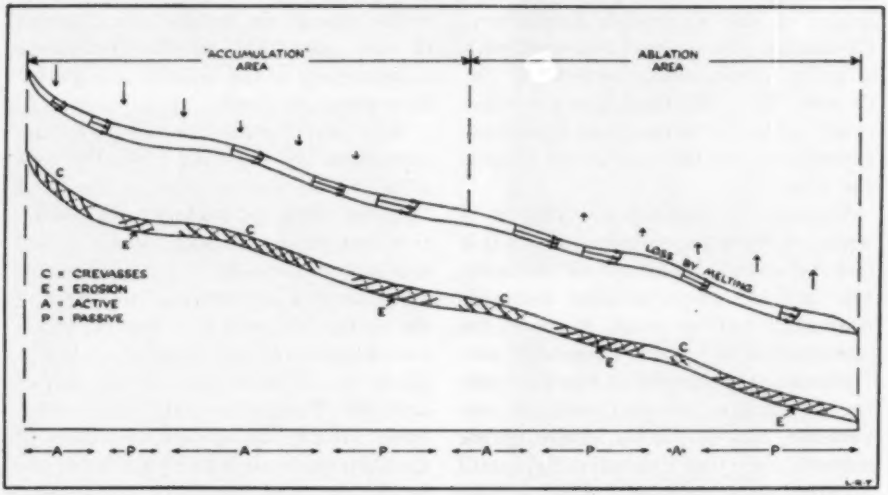


FIG. 1.—Velocity distribution and surfaces of maximum shear stress within an idealized glacier. After Nye

A possible inference from Nye's theory is that in a cirque basin, if rising shear surfaces of the passive state occur only a small distance down-glacier from the plunging shear surface of the active state, something akin to rotational slipping may occur. A stretch of a glacier in the active state cannot, however, be immediately adjacent to a stretch in the passive state, as this would involve two mutually exclusive states in the ice at

the zone of contact. Nye's shear surfaces pass *either* from the surface to the bed *or* from the bed to the surface. If a single arcuate shear surface extends from the glacier surface down to touch the bed tangentially and then continues up to the surface again, then this must represent a rotational slip. Nye considers that a rotational slip may form the limiting case of his theory, but he did not set out to solve the mechanics of the rotational-slip type of shearing—now well accepted in soil mechanics, although its theoretical basis has not been fully worked out.

The passive type of shearing results, at least in part, from the tendency of ice moving relatively quickly down a steep slope to override the slower-moving stretch of the glacier occupying the gentler slope next downstream, rather than to push it bodily forward at the same rate. It seems natural for shear and thrust surfaces to form along the relatively short, curved route to the glacier surface rather than along the greater length of the surfaces of contact between ice and bedrock. Both may, of course, occur together. This overriding motion also serves to thicken the ice occupying the relatively gently sloping part of the bed.

Near the end of a glacier the more actively moving ice of the body of the glacier tends to override the inert ice of the snout itself. This latter ice resists forward motion not so much because of a reduction of bed gradient as because thinning reduces the downhill component of gravity to below that necessary for the ice to be self-propelling. The presence of much included debris may increase the yield stress in the basal layers of this terminal prism of ice. The presence of terminal moraines may also further impede motion.

The upper section in figure 1 shows the theoretical distribution of velocity with depth, assuming an ideally plastic material. In practice one would not expect shearing to occur exactly along the surfaces of maximum shear stress but along the planes of weakness most nearly corresponding with these surfaces. If such shearing and overthrusting movements produce ice bands, these may, in the course of their journey down-glacier, assume different angles to the surface of the glacier. Thus, in examining the angle of intersection between such ice bands and the surface of the glacier, due regard must be paid to their age.

Nye's theory opens up a most hopeful line of investigation for the geomorphologist interested in glacial erosion. If substantial differential movements, in fact, develop along the surfaces of maximum shear stress of the passive state, then erosion may be severe near the points marked *E* in figure 1. The reasons for this are twofold. First, the relatively high velocity in the ice moving down the steep stretch immediately up-glacier may be continued for a short distance at the foot of the step. This may result in backward retreat of the step, together with localized bed erosion at its foot. Second, erosion may be reduced a little farther down-glacier, because shearing, which was largely confined to the basal layers of ice immediately at the foot of the step, is thereafter distributed along both the basal layers and the cycloidal shear surfaces rising toward the glacier's surface. Some of the debris which is always concentrated in the basal ice within a few inches of the rock bed presumably rides up into the glacier along a shear surface and so is no longer available for grinding away the bed. It is for these reasons that we assume that erosion may be relatively severe at the foot of steep slopes and at the ends of glaciers, where shear surfaces of the passive state also

tend to form. De Martonne pointed out many years ago that the down-cutting of the great Alpine glaciers seems to have been most active at the foot of steep slopes in the valley long profiles. In Britain, the depths of glacial lakes, such as Llyn Llydaw, and of the tarns of the Lake District, which have been sounded by students and boys from Brathay Hall at the instigation of members of the Cambridge Geography School, confirm this impression of deep glacial erosion at the foot of steep slopes and cliffs.

This may form an important part, at least, of the explanation of the ability of glaciers to erode basins at the three specific points along their course, i.e., in the cirque, at the foot of valley steps, and near the ends of ancient glaciers, where lakes most frequently occur after the ice has left a region. This explanation is based on far sounder physical principles than is that put forward by Matthes (1949) as a result of the highly original work of Demorest (Flint and Demorest, 1942), whose tragic death at so early an age we all deplore.

PREVIOUS REFERENCES TO ROTATIONAL MOVEMENT IN GLACIERS

Previous references to rotational movement in glaciers have been few, and, as with so many important aspects of glacier mechanics, first mention of this concept is found in the work of Forbes (1842). In this case, however, Forbes was the somewhat cautious recipient of ideas from others. Whewell drew his attention to an article by Milward likening the movement of ice, below an icefall, to that of a rotational slip in an earth or mud slide. Forbes agreed that the sections of a landslip recall the long profile of glaciers, in the matter both of crevasses at the top of the slip and of the upward and forward movement at the bottom. He did not pursue the likeness so far as to suggest any similarity in the mechanics of the movement of glaciers and landslips, but he did attribute the steep up-glacier dip of the ice bands—of which he was well aware—to the resistance to forward motion offered by the great length of the glacier downstream. Further, he thought that these ice bands were induced by shearing and overriding along curved surfaces which intersected the surface of the glaciers at approximately 45°, as Orowan and Nye now suggest. In this he showed remarkable intuition, in view of the ignorance of the physicists of his day concerning plasticity. Forbes's claim that the ice bands were flow bands, i.e., metamorphic in origin, was hotly contested by his friend and rival, Agassiz (1840), who believed them to be bedding bands.

Both T. C. and R. T. Chamberlin were familiar with the emergence, near the ends of glaciers, of inclined ice bands and thrust planes, and R. T. Chamberlin (1928) made the first direct measurement of an overthrust movement. In spite of their signal contribution to glacier mechanics, they did not, as far as we are aware, emphasize any form of rotational movement.

The first clear reference to a rotational movement in a glacier is in a valuable and original paper by Gibson and Dyson (1939). Through an oversight, we have only recently read this paper, and this acknowledgment, though tardy, is none the less genuine. Gibson and Dyson mapped the ice bands of the lower part of the Grinnell Glacier, Glacier National Park, Montana. They gave good reasons for assuming that the ice bands they examined were bedding bands. The outcrops formed sweeping curves across the glacier parallel with the head wall. At

the firn line they dipped up-glacier at 12°, and nearer the snout the dip reached a maximum of 45°. The dip of each band remained constant, as far as they could be traced, to depths of about 40 feet beneath the surface. The bands were about 4–6 feet apart, representing the equivalent of 40–60 feet of snow. This rather large accumulation for an area with an annual precipitation of 30 inches was accounted for by the tendency for snow to collect as a wedge-shaped mass in the sheltered angle where the head wall lower tongue of the glacier was assumed to move as a unit.

This is an ingenious attempt to explain the phenomena which the authors noted; it goes a long way toward the explanation which Lewis has put forward tentatively, but perhaps it does not go far enough. Figure 2 shows some inconsistency in Gibson and Dyson's explanation. The wedge-shaped additions seem to lose their wedge shape when they reach the flat floor. Had all of them been drawn with the correct wedge shape, the

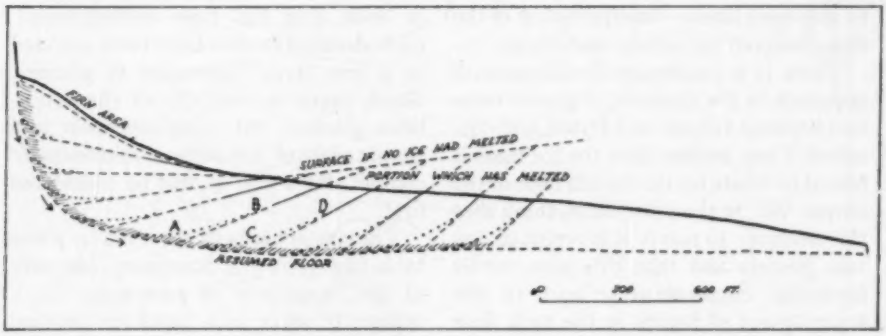


FIG. 2.—Ideal long section through Grinnell Glacier, after Gibson and Dyson, but with dotted curves added.

meets the uppermost limits of the névé. They assumed that the weight of these wedge-shaped additions (fig. 2) caused the mass to rotate because of the curvature of the lower part of the head wall where it flattened out into the cirque floor. The authors thought that this rotation would cease at the point where the floor flattened out, as in figure 2, but that if the glacier occupied a basin, the rotation would continue slightly to the lip. They further considered that the debris that rose to the surface in or between ice bands did so impassively as a result of melting and that little or no overthrusting occurred. Some slight adjustment was envisaged between ice bands. The dip down-glacier would have had to be drawn steeper and steeper until it became vertical, beyond which it would dip steeply down-glacier. Now this is contrary to what was actually observed on Grinnell Glacier, and it is certainly contrary to our own measurements on Norwegian and other glaciers and to all reported angles of dip of ice bands near the ends of glaciers.

Perhaps the most significant difference between the arrangement of ice bands in figure 2 and the arrangement of those which we have examined in the Jotunheim (pl. 1, *A*) is that, in figure 2, the dip continues unchanged to the bed of the glacier, whereas we believe that the

bands curve gently back to join the cirque floor tangentially. Even if they originated as straight-sided bedding bands, they would almost certainly be dragged back by bottom friction so as to curve roughly, as shown by the dotted lines which we have added to Gibson and Dyson's original figure. Then, if we also assume that shearing occurs when the curved bedding surfaces reach a zone approximating to *ABCD* but that farther down-glacier the movement, apart from continued bed drag, is that of a more or less rigid block, we have an alternative—and to us a more likely—interpretation of the dips observed by Gibson and Dyson.

There is a fundamental difference of approach to the question of glacier rotation between Gibson and Dyson and ourselves. They assume that the ice mass is forced to rotate by the curved floor of the cirque. We, on the other hand, think that the tendency to rotate is inherent in certain glaciers and that this may, under favorable circumstances, lead to the scouring-out of basins in the rock floor beneath. True, once such a basin is formed, the tendency for rotation to occur is much increased.

It is impossible to accept fully a hypothesis in which subglacial conditions are assumed which have not yet been directly observed. However, we plan, during the summer of 1951, to tunnel deeply into the small cirque glacier of Skauthöe in the Jotunheim. We hope to reach the head wall at a point far below the bottom of either *randkluft* or bergschrund. It should then be possible to make careful observations of any overriding along ice bands or of any associated rotation of the glacier. Meanwhile, only indirect evidence can be presented in this paper. Yet, even though direct proof of rotation cannot be given, we hope to help in completing the picture of the characteristic features of a glacier so well begun by Gibson and Dyson. These authorities are, as far as we know, among the first to return to the constructive mode of attack on glacial problems initiated by Forbes nearly a century ago.

OBSERVATION ON SKAUTHÖE CIRQUE GLACIER, MID-JOTUNHEIM

Ice bands in glaciers have been studied intermittently by Lewis since 1937, and those in the Jotunheim glaciers have been studied systematically by Clark since 1948 (Clark, 1951). Although a fairly wide area has been reconnoitered, more detailed studies have been confined to a few "type" examples of glaciers. Bands occur in virtually all the Jotunheim glaciers, but Skauthöebreein is a simple slab of ice with a minimum of complications and so will be considered first.

The cirque has been surveyed by plane table (fig. 3) by J. I. Stansbury, formerly of the department of geography, Cambridge, to serve as a basis for detailed work. Plate 1, *A*, taken in late August, 1949, shows a foreshortened view of the glaciers in which the ice bands, together with associated debris, can be clearly distinguished. It is a small relict ice mass lying against the precipitous head wall of its cirque. The surface slope averages 26°, and one simple system of ice bands is exhibited. The transition zone between the banded ice and the névé is usually obscured by a crust of frozen meltwater. The unusually hot summer of 1947, which followed a cold winter with little snowfall, caused quite excessive ablation. Several Norwegian glaciers receded 30 meters a month in the course of the summer, and we were able to examine features which had been previously obscured by snows for many centuries. Streaky, dirt-stained bands which zig-

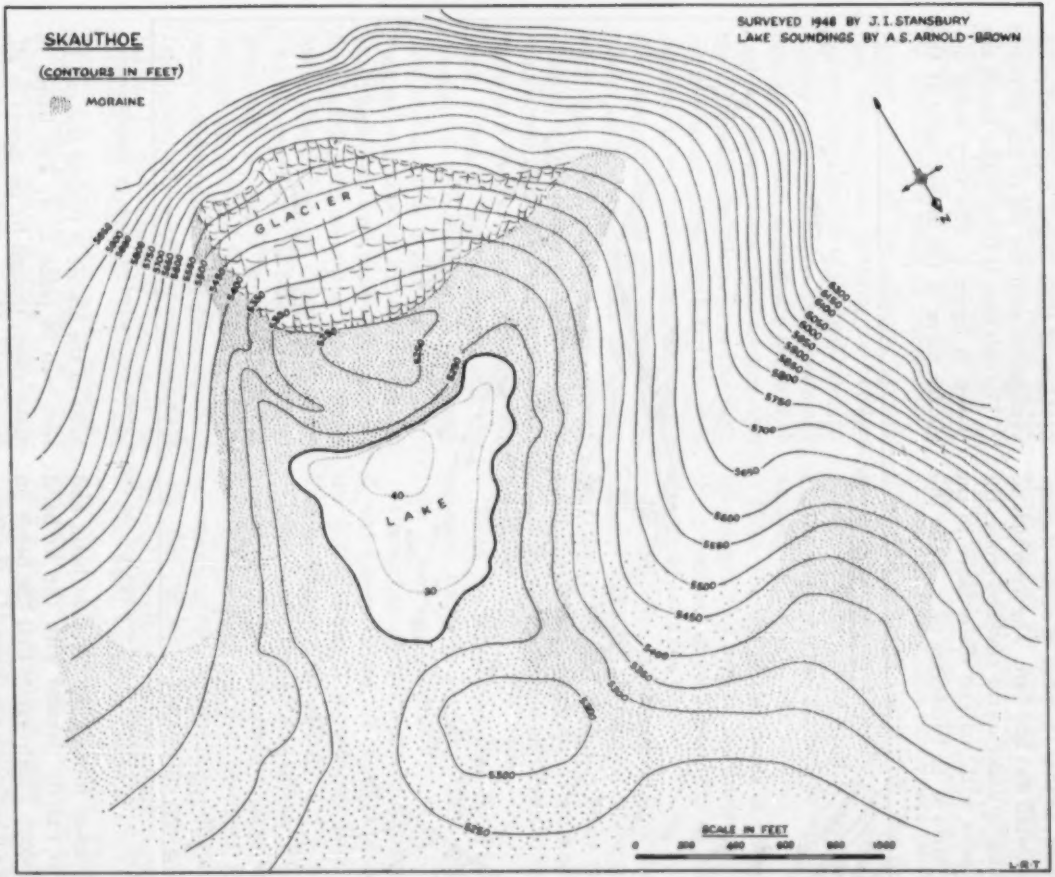


FIG. 3.—Skauthöe cirque and glacier, Mid-Jotunheim, from a plane-table survey

zagged across the névé probably marked the lower limit of the previous winter's snow, so that the névé of earlier seasons' snowfall was laid bare. The ice which ordinarily underlies the névé and crosses the glacier with marked discordance to the system of ice bands (see pl. 1, *A*) on this ly impermeable ice beneath. The angle of dip of these topmost bands and the problem of their origin need further investigation with the aid of dug sections.

The outcrops of three major bands are shown in figure 4. The very sharp upglacier curve of the lowest band at the

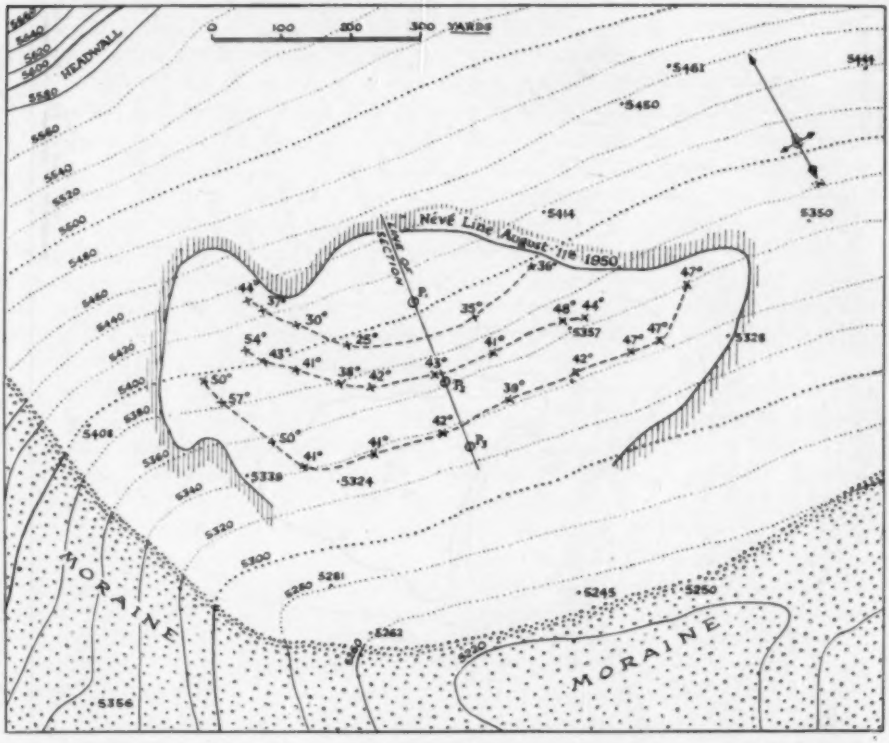


FIG. 4.—The pattern of the outcrops and the dips of three well-marked ice bands on Skauthöebreein

occasion formed the topmost band and was exactly parallel with the ice-band system. This suggests that the bands on Skauthöebreein are formed by the freezing of the soggy water-logged layer at the bottom of the névé. The water in this layer is derived from meltwater which has percolated downward through the névé until it has reached the relative-

sides of the glacier is typical and can be seen well in plate 1, *A*. The bands in many cases, though by no means in all, form little ridges which may protrude, commonly asymmetrically, as much as 6 inches above the furrows on either side. This results from (*a*) the overriding of a band by the one above it and (*b*) differential melting. But some bands pro-

duce no such surface irregularities, or do so only near the center of the glacier. The surface is smoother near the margins and among the uppermost visible bands. When the bands are traced below the surface by digging, they are found to consist of clear, relatively bubble-free ice. Sometimes a plane of discontinuity, against which all the adjacent crystals end, occurs in the center of the band or between the clear ice of the band and the bubbly ice on either side. It is commonly possible to insert a penknife blade with ease into these cracks. They are rare near the margins of the glacier or in the highest and lowest sets of bands but are very common in the central area, although a few narrower bands are without them. These surface irregularities are absent when the discontinuities are absent. Melting often produces surface conditions in which the ridge and furrow structures and the cracks are the most obvious forms at close quarters, and the association with banding is not very clear unless a hole is dug.

Observations of the dip of ice bands on Skauthöebreein were started in 1948. At first, small holes were dug across major bands along a profile down the glacier. The dip could then be read directly and accurately. However, it was found that variations between succeeding bands were sufficient to obscure the main pattern when only scattered readings were taken. It was therefore necessary to measure the inclination of every band along each profile; hence a less laborious method was adopted. A blade or ice-ax was inserted into each crack, and the inclination was read. Trials were made with different observers along the same profile. The greatest discrepancy in observation was 2°, and the average discrepancy was 1°. This is less than the average variation between succeeding bands and so is of sufficient accuracy for our purpose.

Measurements of dip have been taken along profiles up- and down-glacier and also at intervals along individual bands. The results of a complete series from the névé to the snout of Skauthöebreein are shown in figure 5. The profile was taken down the center of the glacier. The following features are typical not only of Skauthöebreein but of band systems in all the other cirque glaciers so far examined.

First, the dip of the bands is always up-glacier. This may be contrasted with the presumed down-glacier dip of the stratification in the névé.

Second, the dip increases from between 0° and 13° at the top to a maximum near the center and then dies away toward the snout, commonly to an angle of less than 10° and in many cases to the horizontal.

Third, the difference in the dip of two adjacent bands is in many cases more than 2° and may exceed 10°.

These characteristics appear in figure 6, which summarizes the dip data of figure 5 and shows it superimposed on a long profile down the glacier.

Dips vary considerably along individual bands. Several series of measurements have been made at 10-meter intervals along the outcrops of such bands. The dip reaches a maximum near the sides of the glacier on the upturned part of each band, though not right at the top. The dip is least near the center. There are again considerable local variations. Such a series is shown in figure 7. Also, results of dip measurements at various points along three major bands are shown on the map of outcrops (fig. 4).

The most significant feature of the evidence here presented is the remarkable generalized curve followed by the

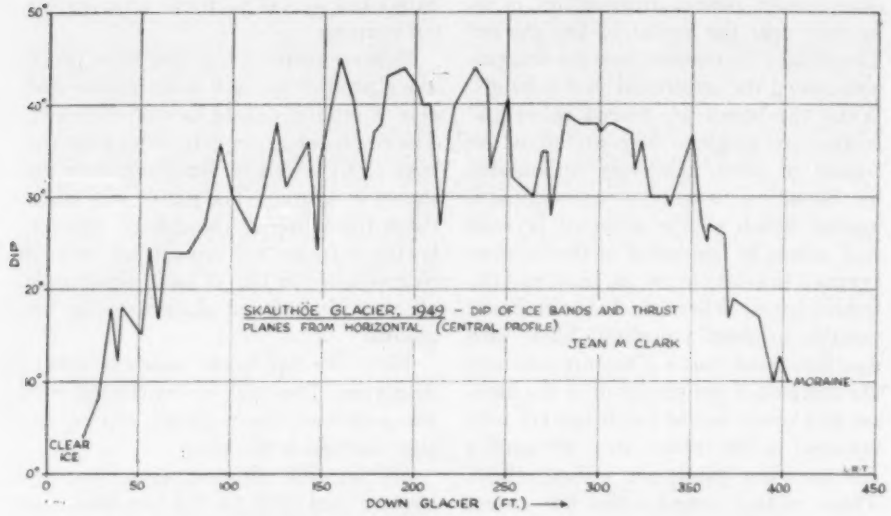


FIG. 5.—Dip of ice bands and thrust planes down the middle of Skauthöebreein, 1949

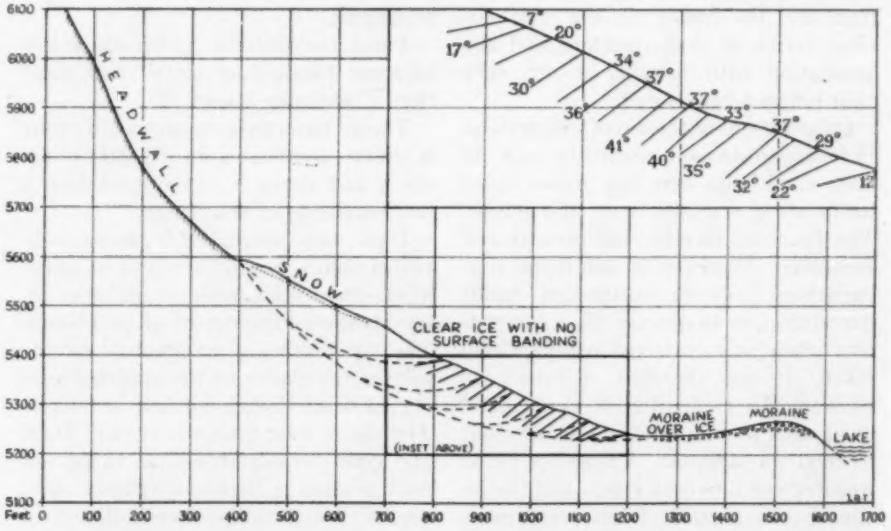


FIG. 6.—Long profile down Skauthöebreein, showing outcrops of ice bands and thrust planes and average dips of groups of four ice bands, 1949. Values for dips based on average for groups of four. Broken lines represent assumed rock bed and shear plane.

plot of the dips of 65 ice bands down the central profile of Skauthöebreein (fig. 5). There must surely be some fundamental reason for this systematic increase in the dip from the névé to midway down the glacier and for the rather more localized decrease in dip as the end moraine is approached. Figure 6 more clearly relates this arrangement of the dip to position on the glacier. The interpretation already suggested by Lewis, though it can hardly be presumed to be correct in detail, does seem to afford a fairly satisfactory explanation of the changing dip of the bands. Lewis assumed that the weight of the new snows each season, together with the melting on the lower half of the glacier, upset the equilibrium in the upper half of the glacier until a rotational slip—or a Nye-type slip—occurred. This is represented in position approximately by the uppermost curving broken line. Movement may occur along several neighboring surfaces, so that the differential movement along any particular one may be small. A prerequisite for the successive down-glacier increase in tilt of the band is basal slip more or less following the rock bed, thus causing the whole glacier to rotate. Continued melting near the terminus, while assisting the rotation, would also bring the deeper

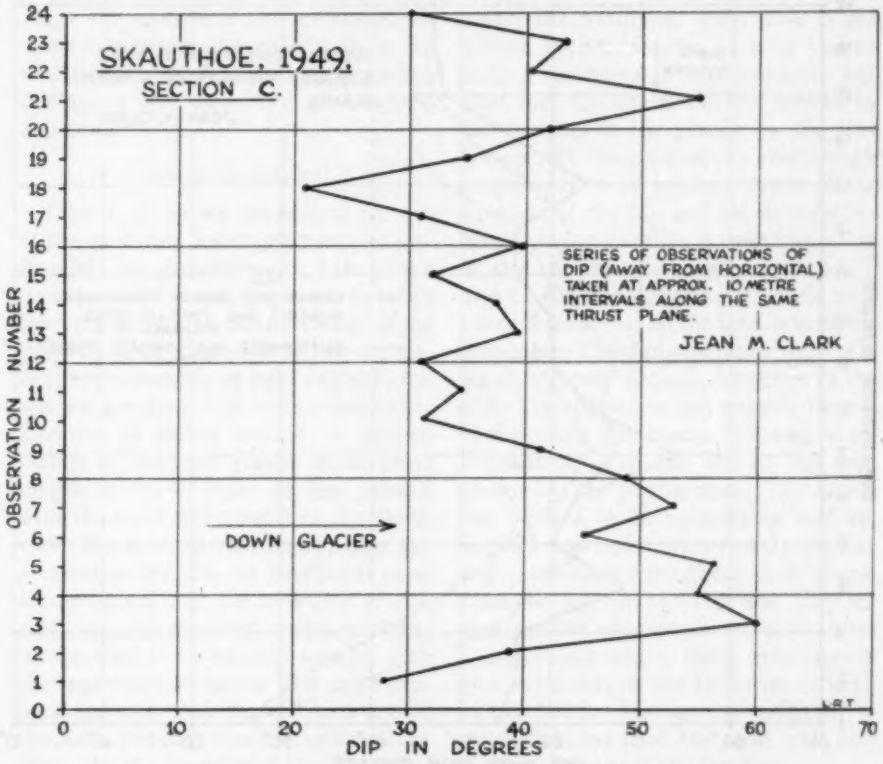

SKAUTHOE, 1949.
SECTION C.
SERIES OF OBSERVATIONS OF DIP (AWAY FROM HORIZONTAL) TAKEN AT APPROX. 10 METRE INTERVALS ALONG THE SAME THRUST PLANE.
JEAN M. CLARK
DOWN GLACIER
OBSERVATION NUMBER
DIP IN DEGREES
0 1 2 3 4 5 6 7 8 9 10 11 12 13 14 15 16 17 18 19 20 21 22 23 24
0 10 20 30 40 50 60 70

FIG. 7.—Dips at 10-meter intervals along a single thrust plane, Skauthöebreein

parts of the glacier to the surface. In these deeper parts, the shear or overthrust surfaces within the glacier depart only slightly from the direction of the glacier bed, which they join tangentially. Therefore, when these deeper layers rise to the surface, the dips of the overthrusts and bands become reduced. It must not be expected, of course, that, once an internal shear or overthrust surface has formed, it will move down through the glacier without gradual and perhaps considerable alteration of shape.

There may be other reasons for the increase, then later decrease, of the up-glacier dip of the ice bands and overthrust surfaces, but rotation of the whole glacier would seem to offer the simplest explanation. For a short glacier like Skauthöebreein, this single rotation may suffice to explain the dips, though further independent rotation in the upper half of the glacier may also occur.

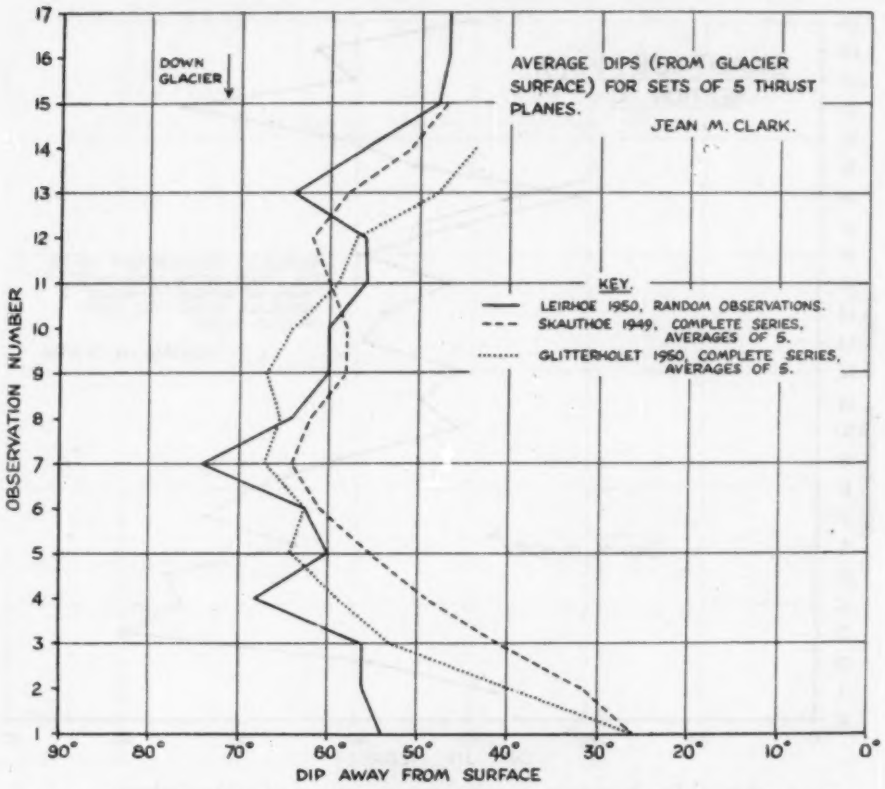


FIG. 8.—Average dips of sets of five thrust planes, for Leirhöe, Skauthöe, and Glitterholet glaciers, Mid-Jotunheim.

This arrangement of ice bands in Skauthöebreein seems to be typical of such small, steep Norwegian glaciers. A cursory examination of Smedbotn Glacier in the Rondane and closer examina-

tion of the similar glacier occupying Glitterholet on the west side of Glittertind, Mid-Jotunheim, clearly show similar features, as figure 8 illustrates. We hope to gain more definite indications of the movement of Skauthöebreein and of the source of the great boulders that reach the surface in the ice bands, always so carefully aligned as to lie in the plane of the bands, after our tunneling operations and associated observations this summer.

JUVASS KJEDLAN

Plate 1, *B*, shows the general pattern of the ice bands outcropping on the surface of the Juvassbreein. The cirque faces east, and the photograph was taken from the cliff top on the north side of the cirque, looking diagonally across the glacier, approximately at right angles to the steepest gradient of its surface and to the direction of glacier motion. A marked feature of the band system is the great up-glacier curve more or less parallel with the main embayment of the headwall. The head of this embayment approximates the zone of maximum snow accumulation, and the névé line of August, 1949, after a very snowy summer, can be seen to be roughly parallel with the uppermost ice bands. This again supports the view that the ice bands are, in the main, bedding bands of sedimentary origin. In the foreground the bands recurve up-glacier and disappear under the fresh firn, making a marked angle with the firn line on this side of the glacier. This backward curve probably represents the retardation of the marginal ice by frictional drag. The arrangement of the straight crevasses in the foreground also supports the suggestion that the central ice moves faster than that near the sides and so gives rise to the usual diagonal arrangement of the tension crevasses. The side drag may be accentuated by the slower movement of the thinner ice on this "warmer" side of the glacier.

The general movement of the glacier today is diagonally across the cirque toward the northeast moraines and away from the zone of maximum accumulation under the southwest cliffs. This is fortunate for our studies, because the resulting movement is approximately parallel with the line of the cliff formed by the melting of the glacier by the lake waters. We therefore have a ready-made section cut so as to show much of the progress of the firn and ice through the glacier to the moraine. A close inspection of this cliff face, as shown in plate 1, *B* and *C*, the photographs for which were taken from a boat on the lake, is most informative. These photographs were taken after Lewis' (1949*b*) reference to the cliff. They illustrate and amplify several of the points then made. The zone of accumulation is to the left of the area shown by the photographs. The bands can be seen to be remarkably uniform in thickness and extraordinarily regular and continuous throughout their course from cliff top to the water line. This argues against any simple evolution from a sedimentary origin, which would surely give variations in the thickness of individual bands. The annual addition is usually thicker near the head wall and wedges out toward the firn line.

The dip of the bands at the top of the cliff can be seen to increase to a maximum just right of the center of the photograph, from which point they follow sweeping parabolic curves, arcuate, at first, but straightening as the dip becomes more gentle near the foot of the cliff. When photographed in 1946, these ice bands, indicated by included debris, were more arcuate throughout their length. Lewis' suggestion that they may

correspond with shear surfaces, either of the rotational-slip type or of the Nye type, would be somewhat more convincing if their courses were roughly arcuate, at least in the zone where they were formed. Apparent differences in their form from year to year may be due to their being "active" only in favorable seasons, after which normal glacier flow distorts them. One feature clearly shown in the right of the photograph strongly supports this partial "metamorphic" origin for the bands. The steeply dipping and markedly curved bands overlie others which have a more gentle and uniform dip. In the V-pattern thus formed, several of the upper bands with uniform dips are cut off short. This represents a form of unconformity which cannot be accounted for by differences in the ablation seasons or in the abundance of accumulation in any one year; for the bands within the V thin downward, that is, toward the zone of maximum accumulation near the head wall and not toward the firn line, as bedding bands should.

The most reasonable explanation seems to be that the mass bounded by the lowest curved bands has rotated and thrust over the more uniformly dipping bands beneath. There is usually a slight upward bulge in the surface over this supposedly up-thrusting mass of the glacier. This may, of course, be due to a greater compaction of the ice forming this mass. This feature was also evident in 1939, when a cursory examination of the surface of the glacier suggested no significant difference in the compaction of one part of the glacier compared with the other in this zone; A similar V-arrangement of bands was also visible in the summer of 1950.

If this interpretation of the curved bands overlying and cutting off the uniformly dipping bands is correct, this example also shows how bands adjacent to the shearing surface may adjust themselves to the stresses involved in the movement. No single curved band transects the plane bands; instead, the upper plane bands become curved about halfway down the cliff face, and only the plane bands at the extreme right of plate 1, *C*, remain plane throughout their visible courses. In fact, the whole pattern of the V-banding indicates considerable distortion from their earlier form, whatever their origin may have been.

Plate 1, *D*, shows the remaining portion of the ice cliff as far as the end moraine. The dip is seen to become more and more gentle toward the snout, an apparent steepening of the dip near the terminus being due, at least in part, to a change in the direction of the ice cliff. The nature of the movement of this part of the glacier may well be that normal to glaciers, i.e., with the velocity a maximum at the surface and falling off because of bottom friction as the bed is approached.

A further illustration of the gradual distortion of ice bands by localized stresses is shown by the behavior of the ice bands around the two holes shown in plate 1, *D*. The bands immediately above each hole have sagged down, and those below have risen to help fill the hole. Only the bands in the region of greatest stress adjacent to the holes had yielded plastically, whereas those a little farther away had remained unaltered. We may therefore be justified in assuming that the system of banding shown by the glaciers as a whole has responded to stresses within the ice and may mark zones of maximum stress. It should, however, be clearly borne in mind that, once modified, the band system travels down-glacier away from the zone of metamorphic change. Any given bands may therefore

bear the impress of past strains and not necessarily indicate zones where modifying forces are active at the time of observation.

LEIRHÖEBREEIN

Consideration of one further set of observations is appropriate. Preliminary work on the glacier on the summit plateau of the Leirhöe massif (6,000–7,000 feet) began during the 1950 season. This glacier is larger than either Skauthöe or Juvass and flows westward, out of its cirque, over a step, and on toward the edge of the plateau. Only the southern side wall and the eastern back wall are present, the glacier being bounded on the north by moraine material. The relief is considerably less than in the case of the cirques so far described, the maximum height of the surrounding cliffs being about 250 feet. The step runs north-south, and the rock of the riser can now be seen through a window in the center. Unfortunately, the area bare of snow in 1950 was very limited. A number of interconnected band systems occurred, two originating under the main rock walls and one on each side of the bare rock in the center of the step. Observations of dip were taken along two profiles at right angles to the outcrop of the bands, line *I* (fig. 9) crossing the bands parallel to the side wall below the step, and line *II* crossing the bands over the step to the south of the rock window. The observations were incomplete because of lack of time and the amount of frozen crust present, which concealed the bands beneath.

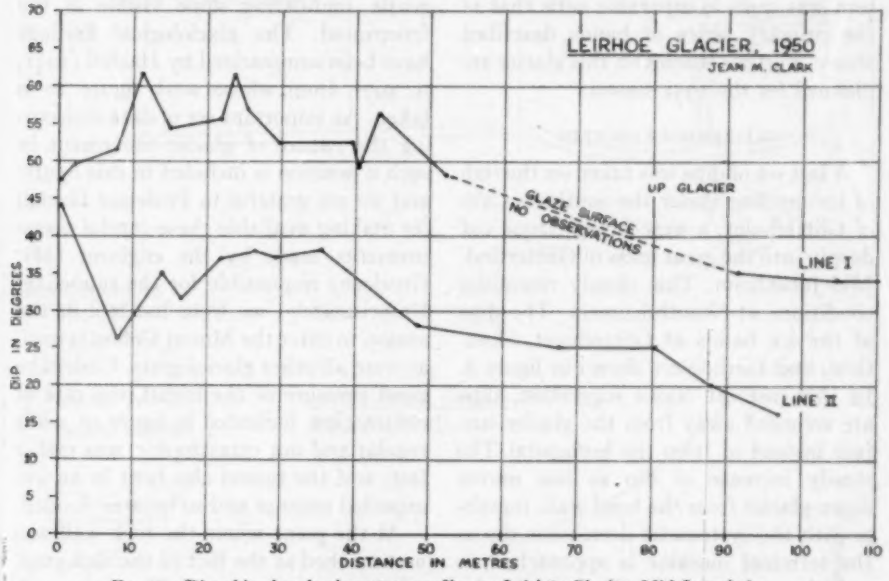


FIG. 9.—Dip of ice bands along two profiles on Leirhöe Glacier, Mid-Jotunheim

The outcrops of the side-wall bands of line *I* were orientated at right angles to the usual one. They were parallel with the steepest surface gradient (17°20′). The slope at right angles to the outcrop was only 6°30′. The distribution pattern given by plotting the observations, how-

ever, is of the usual type. The maximum surface slope of line *II* was 20°30′ at right angles to the outcrops, i.e., following the normal relationship. The dip of bands just above the step was less than 20°. The dip increased slightly over the step and rose to a maximum of 48° and then fell again to 26° near the foot of the step. The last reading, 45°, was apparently anomalous. However, the general pattern was quite comparable with that of the primary series of bands described above. Further studies on this glacier are planned for the 1951 season.

GLITTERHOLETBREEIN

A last set of dips was taken on the slab of ice nestling under the southern walls of Glitterholet, a west-facing cirque cut deeply into the great mass of Glittertind, Mid-Jotunheim. This closely resembles conditions at Skauthöebreein. The dips of the ice bands at Glitterholet, Skauthöe, and Leirhöe are shown in figure 8. In this case, at Nye's suggestion, dips are recorded away from the glacier surface instead of from the horizontal. The steady increase of dip as one moves down-glacier from the head wall, together with the systematic decrease in dip as the terminal moraine is approached, is well shown in all three. The two simpler glaciers show this characteristic particularly well. Whether or not we are correct in attributing this remarkably systematic behavior of the dips of the bands to one or more forms of rotational movement remains to be seen.

THE MOUNT COLLON ICEFALL, AROLLA GLACIER, SWITZERLAND

Plate 2, *A*, shows the 3,000-foot Mount Collon icefall joining the Lower Arolla Glacier. Lewis visited this glacier in July, 1949, at the instigation of Gerald Seligman and through the kindness of Professor Haefeli, scientific adviser to the West Switzerland Electricity Company which wanted to find the position of the rock wall near the foot of the fall. They therefore arranged for a tunnel to be dug through to the back wall. The dump of black debris at the mouth of the tunnel can be seen in the middle of the photograph, where the glacier flattens from its tumultous descent into the relatively gentle, undulating slope visible in the foreground. The glaciological findings have been summarized by Haefeli (1951, p. 499), from whose work figure 10 is taken. An important set of data concerning the nature of glacier movement in such a position is included in this figure, and we are grateful to Professor Haefeli for making available these careful measurements made by the engineer (Mr. Grosjean) responsible for the tunneling. Unfortunately, we were too late in the season to enter the Mount Collon tunnel, as were all other glaciologists. Under the great pressure of the icefall, the rate of contraction, indicated in figure 10, while regular and not catastrophic, was rather fast, and the tunnel also bent in an unexpected manner and so became flooded.

At the point where the rock wall was first reached at the foot of the rock step, a large cave was tapped. The water suddenly released caused a dangerous flood in the tunnel. The velocity and rigidity of the ice in the fall were sufficient to cause it to move slightly away from this rock wall where the wall became vertical, rather as the water at the foot of a great waterfall does. That this could happen nearly 100 meters below the surface of the ice serves as a timely reminder of the strength of ice under pressures giving rise to stresses considerably greater than the yield stress. The ice, in behaving in this manner, showed no signs of acquiring greater plasticity under pressure, as

the supporters of the extrusion-flow hypothesis would have us believe. The rate of yield from November 26 to December 30 at the point marked $A'A'$—about 100 feet below the surface—is also shown in the figure. This shrinking was roughly proportional all around the cross section, the floor rising at much the same rate as the roof collapsed, the diameter becoming reduced about 1 per cent per day.

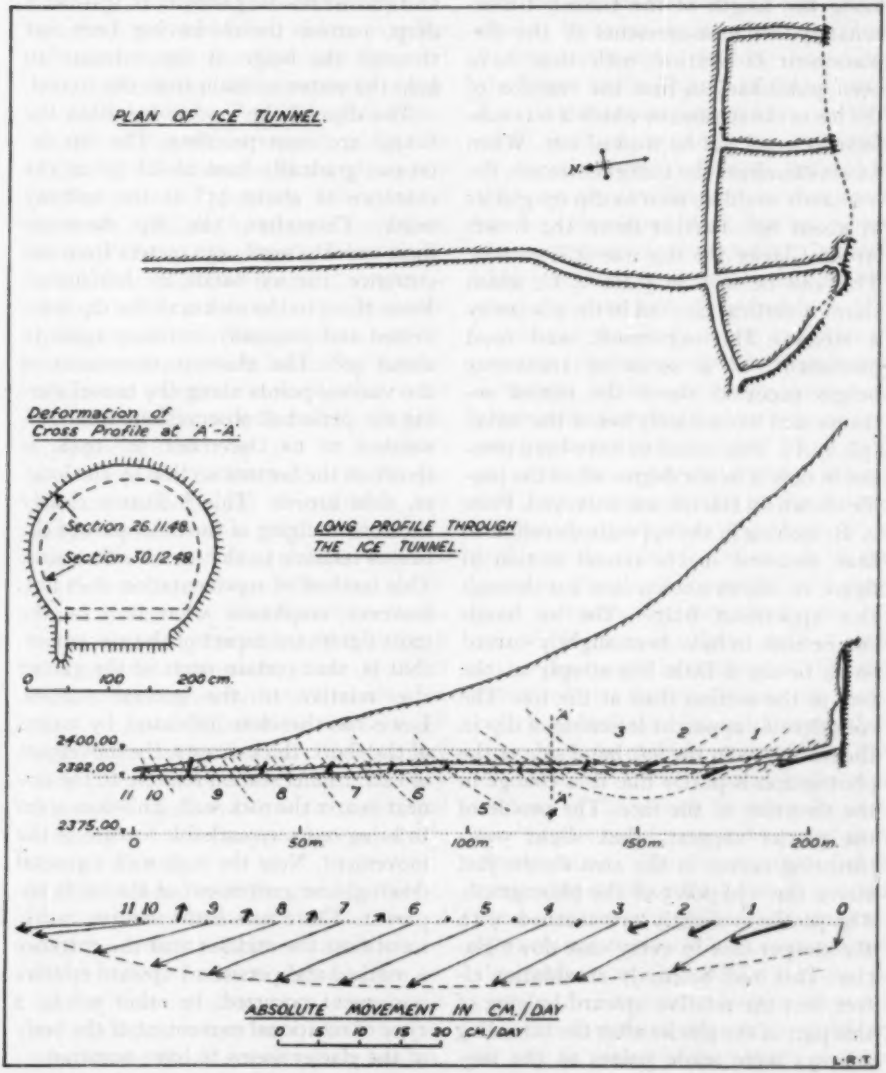


FIG. 10.—Ice bands and glacier movement in the Mount Collon tunnel, Lower Arolla Glacier, Switzerland, 1949.

Among the more important results obtained from the tunneling are indications of the dip of the ice bands from the surface to the bed of the glacier and the absolute movement of the different points along the length of the tunnel. Unfortunately, no measurements of the displacement of vertical with time have been published, so that the reaction of the ice to the stresses to which it was subjected cannot yet be worked out. When Lewis examined the tunnel entrance, the ice bands could be seen to dip up-glacier at about 60°. Farther down the Lower Arolla Glacier the dip was still greater. This can be seen in plate 2, *C*, which shows a vertical face cut in the glacier by a stream. The uppermost, and most prominent, of a series of transverse bulges occurred above the tunnel entrance and immediately below the icefall (pl. 2, *A*). This seems to have been present in only a minor degree when the profile shown by Haefeli was surveyed. Plate 2, *B*, looking in the opposite direction to that assumed in the tunnel section in figure 10, shows a steep face cut through this uppermost bulge. The ice bands can be seen to have been slightly curved so as to dip a little less steeply at the foot of the section than at the top. The considerable apparent lessening of dip in the ice shown in the left-hand side of the photograph is partly due to a change in the direction of the face. The profile of the glacier suggests that slight overthrusting occurs in the area shown just above the mid-point of the photograph. The profile is slightly saw-toothed, with the steeper face in every case down-glacier. This may be partly an ablation effect, but the relative upward bulging of this part of the glacier after the tunneling surveys were made points to the suggestion of overthrusting being a reasonable one.

The plot of absolute movement shown in figure 10 indicates that another relative bulging-up occurred near the entrance during the tunneling operations. This reversed the slope near the entrance and caused flooding within, in spite of a deep, narrow trench having been cut through the bulge at the entrance to help the water to drain from the tunnel.

The dips of the ice bands within the tunnel are most puzzling. The dip decreases gradually from about 70° at the entrance to about 45° at the halfway mark. Thereafter, the dip decreases more quickly until, 150 meters from the entrance, the ice bands are horizontal. From there to the rock wall the dip is reversed and gradually increases again to about 30°. The absolute movement of the various points along the tunnel during the period of observations from November 20 to December 30, 1948, is shown in the bottom section by the longer, thin arrows. This indicates clearly the down-bulging of the inner part of the tunnel relative to the ice on either side. This method of representation does not, however, emphasize what may be the most significant aspect of the movement, that is, that certain parts of the glacier rise relative to the general surface. Lewis has therefore indicated, by means of the short, thick arrows, the movement of each tunnel station relative to the one next nearer the rock wall. This does seem to bring out a remarkable feature of the movement. Near the rock wall a general down-glacier movement of the ice is apparent. There was little relative movement near the middle; near the entrance a marked and persistent upward relative movement occurred. In other words, a type of rotational movement of the body of the glacier seems to have occurred.

The choice of this method of representation was not entirely arbitrary, be-

cause what we really desire to know is the movement of the different parts of the glacier at depth relative to the underlying layers which are carrying them forward. For instance, the absolute movement of the glacier at the tunnel entrance is the sum of all the movements of the underlying ice between that point and the rock bed. It is particularly suggestive that near the rock wall differential movement is roughly along the ice bands. These ice bands may therefore represent active shearing. Down-tunnel from the syncline the relative movement was small and was upward and backward. One might be tempted to ascribe this to minor inaccuracies in the measurements, if this feature were not repeated on a gradually increasing scale at a series of points successively nearer the entrance. As the direction of the relative movement shown by the thick arrows near the tunnel entrance is markedly discordant with the dip of ice bands, one might conclude that the bands in this zone were formed somewhat farther up-glacier, perhaps where differential movement of the type that had formed the uppermost bulge had occurred. Perhaps this upward movement is compounded from one or more rotational shears within the ice—possibly including one immediately adjacent to the rock. If this inference proves to be correct, then we may have the first evidence of rotational shearing of the Nye type occuring at the foot of an icefall.

Our impression of the great complication in glacier movement at the foot of such an icefall is further supported by the presence of several active bulges occurring down-glacier from the Arolla tunnel. These bulges were about 80 meters apart, which, as Grosjean told us, approximated the annual movement of the glacier in that zone. The surface of the glacier was distinctly dirtier in the troughs than on the crests of the waves, and farther down-glacier, where the waves gradually disappeared, the marked alterations of black and light stripes crossing the glacier in great arcs downstream became even clearer. These Forbes bands are a striking feature of the Lower Arolla Glacier, at least in summer. The waves cannot result from differential melting between the relatively clean and the dirty surfaces, because in that case they would have become more strongly developed down-glacier, where the color contrast was greater. Instead, they gradually flattened as they moved away from the foot of the icefall.

Most of the crests of the waves stood at much the same level as the trough next upstream. Some, however, rose well above the surrounding glacier surface, causing a reversal of slope. Interesting examples of antecedent drainage among the little streams on the glacier showed clearly that certain of the ridges were still actively rising in the course of the summer. Plate 2, *D*, shows the third wave crest down-glacier from the tunnel entrance. It rose, in places, fully 3 feet above the trough next above it. The old course of the streams before the wave crest had risen and caused a reversal of slope can be readily recognized. The "consequent" streams spread lazily over the area of the trough, showing distributary patterns and other evidence of slow down-cutting. The consequents fed a transverse stream, which joined a larger consequent stream, which had been powerful enough to maintain its course across the rising wave. This stream is shown in plate 2, *E*. The gorge through the wave and the ill-drained flats next upstream all help to complete the picture of antecedent drainage.

The approximate rate at which little

streams could incise their courses was indicated by the lowering of the thresholds of little pools on the glacier's surface. These pools froze at night, and the stream flowing out of them cut down into the glacier in the heat of day and tended to empty the pools. Where these were partly sheltered from the sun, some of the frozen crust would survive to the following day. The daily fall in level of the pools could then be seen to be about 2 inches. Grosjean informed us that the ablation of the glacier surface was about 5 cm. a a day; and so it seems that small streams lowered their courses at much the same rate as did the general glacier surface. Where upward bulges formed, this delicate balance seems to have been upset, and most of the streams were diverted. The really large ones were, of course, well able to incise their courses 5 or 10 feet into the glacier surface and very much deeper still nearer the snout.

There are probably two main reasons for the waves and troughs gradually being transformed into Forbes bands. First, in the early summer when snow has left the countryside bordering the glacier, some still remains in the troughs, where they are deepest, near the foot of the ice-fall. This snow, which is already probably fairly deep, traps more dust than does the purer ice exposed on the crests of the waves. Further, the pools and soggy patches occurring in the troughs right through the summer trap both dust blowing about and also silt brought down by streams and runnels. Thus dirt is concentrated in the troughs, and, once it has collected there, no further agency redistributes it. So the alternate bands move steadily down the glacier, acting as most convenient indicators of the distance moved by the glacier in a year.

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DEVELOPMENT OF END MORAINES IN EAST-CENTRAL BAFFIN ISLAND[1]

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ABSTRACT

Five steps in the accumulation and deposition of end moraines are postulated from observations of profiles and ablation features around the south end of Barnes Ice Cap. This end of the icecap is generally more expanded than it was throughout the past century, but a retreat, producing end moraines, has begun on the southern and southwestern sides.

Dirt is raised to the surface on shear planes along a narrow fringe of black ice around the edge of the icecap. This is till, exposed in small amounts by 5–7 feet of ablation each summer. This film of debris slips and slides down the 10°–30° marginal slope and gathers over the lower ice slope. Where the ice becomes blanketed by 3 feet of ablation moraine and dirt-filmed ice above melts back rapidly, a trough develops parallel to the margin, isolating a steep-sided moraine on an ice core. After many years this material is let down irregularly to solid ground by flow of the till cover, cuts by transecting streams, and lateral undermining of the exposed core ice.

INTRODUCTION

The complete accumulation and emplacement of an end moraine involves more than one summer season. It was obvious at the outset of the Baffin Island Expedition of 1950[2] that the whole event could be neither witnessed nor measured at any one spot. The best substitute consisted of (1) careful observation of processes operating along a retreating ice edge and (2) comparison of different sectors of the ice edge, in order that stages might be deduced. Some 40 miles of the southeastern margin of Barnes Ice Cap were inspected on foot or ski; another 75 miles were observed from the air (fig. 1). From these observations the production of the end moraine is pieced together.

It is not proposed that all end moraines form this way. The Barnes Ice Cap is 90 miles long and 45 miles wide; hence its climatic regime cannot be identical with that of a large Pleistocene ice sheet. Furthermore, it is a cold, arctic icecap: the deep ice proved to be about 13° F., which is undoubtedly near mean annual temperature.[3] Such ice is more viscous than ice near 32° F. (Dorsey, 1940, p. 455), and it cannot contain deep subglacial streams as warmer ice does. Nevertheless, the features produced here are similar to end moraines in many parts of North America, and it is hoped that the study of moraine formation here will shed light on that of the Pleistocene (Flint, 1947, pp. 127–130).

PAST EXPANSION AND INITIAL RETREAT

The south end of Barnes Ice Cap at 69°33′ N. to 69°55′ N. on north-central Baffin Island is as large today as it has been for several centuries. This is shown strikingly by the plants. On coarse soils right up to the ice edge thick tufts of *Cassiope* (*C. tetragona*) and mosses are established. Expedition botanists and others advise that this takes at least 25 years or even a century to become established.[4] Several species of lichen grow

[1] Manuscript received June 2, 1951.

[2] General account by Baird (1950), leader of the expedition.

[3] Mean annual temperature at Clyde, 100 miles away on the seacoast, is 11°.1 F.

[4] P. Dansereau was in charge of botanical studies, and M. E. Hale was lichenologist. Both men visited the Barnes Ice Cap briefly and expressed opinions verbally.

profusely on bouldery areas right up to the moraine of the active glacier. The lichenologist estimated 50 years as a minimum time for these dense large colonies to develop.

Two geological features indicate some expansion in the recent past. One is the great disparity between past and present directions of ice motion at the very southeastern edge of the icecap. It is fair to assume that motion today is radially southward and southeastward from the high central dome (at 3,700 feet altitude) to the low ice edge (below 2,000 feet). Yet, immediately adjacent to the southeastern margin, there are *roches moutonnées*, striae, and lunoid furrows, expressing northeastward motion parallel to the ice edge (fig. 1). They were made by ice with a very different center of motion. As centers of ice accumulation must shift slowly, these features were formed probably several centuries ago. Second, the southern third of the ice margin is immediately adjacent to fully formed, sorted polygons and stripes of boulders as

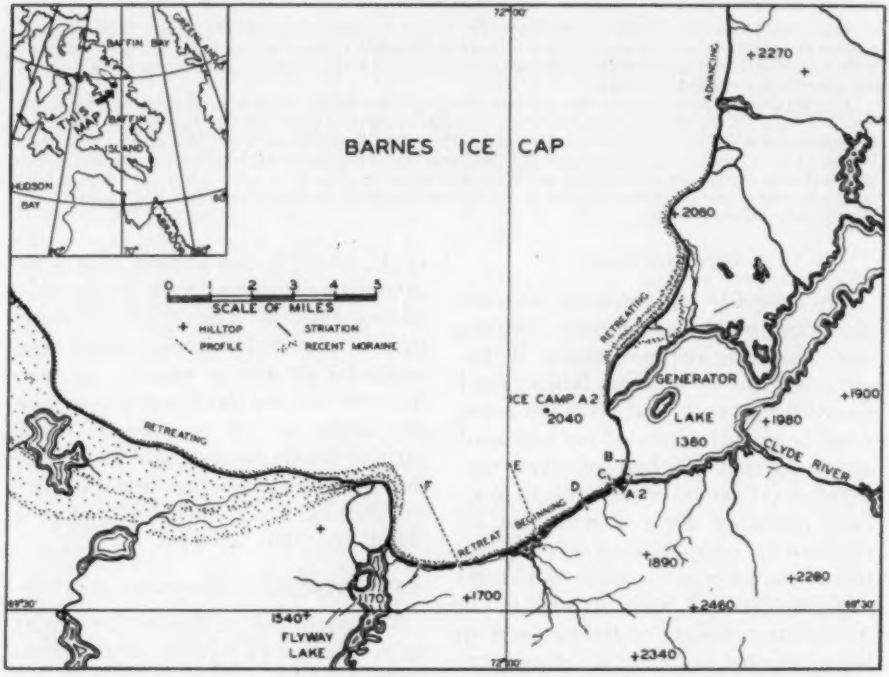

FIG. 1.—Sketchmap of the southern end of Barnes Ice Cap, showing the steep glacier margin, areas of accumulating moraine, and altitudes of hills (Paulin).

PLATE 1

A, Belts of recent retreatal moraine along the southwestern side of Barnes Ice Cap.
B, Shear planes filled with dirt, just east of profile *D* at the southern end of Barnes Ice Cap.

A

B

Retreatal moraine belts and dirt-filled shear planes, Barnes Ice Cap

A

B

Black ice slope and rock table

well as fissure polygons. Granted that some boulders may be gathered into miscellaneous groups rapidly, recent studies in Alaska[5] and observation of contrasts on fossil and active outwash trains on Baffin Island indicate that patterned soils and even large ice wedges take one or more centuries to form. These many areas of patterned soil, identical in form to those in Alaska, could not have been disturbed by glacier movement for many decades.

In addition, there is climatological evidence of expansion along the northeastern side today. The inclination of the outer 500 feet of the icecap averages 10°–20°, and the sharp angle between ice edge and adjacent land catches deep snowdrifts in the winter. These did not completely melt in the summer of 1950; it was estimated that half the northeastern ice edge was still covered by snow in late August. Air photographs taken in July, 1948, show some drifts for that year.[6] Thus there is some net addition of snow, which turns to ice along the very edge; therefore, Barnes Ice Cap is growing here and there by marginal accretion.

This general expansion has already turned to initial retreat on the south and southwest. On these sides the outer 500 feet of icecap generally slope only 5°–10°. For 20 miles there are three belts of retreatal moraine parallel to active ice (pl. 1, *A*). These are not yet fully inhabited by local tundra plants, and annual rings in dwarf willow (*Salix arctica*) on the outer moraine show a maximum of 25 years.[7] At the southestern end of this belt the inner two moraines join in one area 2,500 feet broad, in which the ice is heavily covered with ablation moraine. Comparison of the photographs taken in the summer of 1948 with the ice margin as it was in 1950 suggests thinning and additional dirt cover. Probably this retreat set in locally on the southern and southwestern sides a few years ago. The net effect of a general recession on the southwest and advance on the northeast is a very slow northeastward shift of this end of the icecap.

[5] Personal communication with R. F. Black, U.S. Geological Survey, January, 1951.

SOURCE OF MORAINAL DEBRIS

All the debris which makes up the present superglacial moraine came up on shear planes through the thin edge of the glacier from the rock bottom. All ice up to 100 or 200 feet above the steep toe is layered with myriads of dirt-filled thin fractures, striking roughly parallel to the ice edge (pl. 1, *B*). The dirt filling in each shear crack is from $\frac{1}{100}$ inch to 1 inch thick; generally it is less than $\frac{1}{10}$ inch, although dirt is smeared widely at the exposure. Small pebbles and even large boulders are seen occasionally where they were in transit up these shear planes. Shear planes above and below such boulders and till clots swell around them in "augen" structures. Some of these bands crop out horizontally along the steep ice

[6] Much use was made of trimetrogon photographs flown by the Royal Canadian Air Force in 1948 and 1949.

[7] Other plants collected indicate brief stable conditions but no great age: *Cerastium alpinum*, *Luzula nivalis*, *Oxyria* (sp. ?), *Saxifraga oppositifolia*, *Stellaria longipes*.

PLATE 2

A, Upper edge of the black-ice slope, showing veneer of creeping dirt on the right. The uppermost dirt-laden shear crack overhangs by shove from the main glacier on the left.

B, Rock table consisting of a 4-foot boulder on a 15-inch-high pedestal of ice retaining original shear planes. Near profile *E*.

edge for thousands of feet, but most of them die out in 10–100 feet. The clear ice layer between seams varies between $\frac{1}{8}$ and 6 inches in thickness. Multitudes of meltwater channels, 3–15 feet deep, transect these shear planes, revealing that they dip 10°–36° back under the main icecap (fig. 2).

It is apparent that the layers of ice between shear planes move very slowly over one another, like a deck of cards pushed up an incline. The 30° dip back under thicker ice corresponds to faults by this increase of motion higher in the ice, however; so it may be that motion in any one year is dominantly along a selected few shear planes here and there, while intermediate fractures suffer no displacement.

Dirt filling in the shear planes is restricted to the very edge of the icecap. On the air photographs and in measured profiles the average width of the dirt-smeared surface ice is 450 feet, which is about 1 per cent of the whole area of active icecap. The upper limit of this dirty

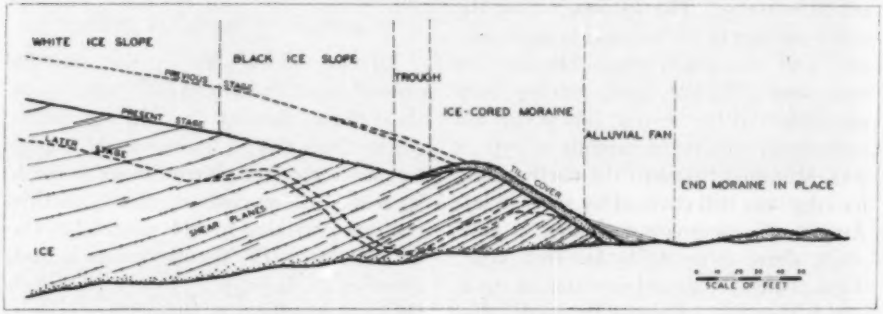

FIG. 2.—Diagram of the retreating margin of Barnes Ice Cap, based upon averages measured. Dots indicate dirt in transport up inclined shear planes, or gathering on the surface as ablation moraine, and finally deposited as end moraine (*right*).

mass thrust in rocks. On one ice cliff at the extreme southeastern end of Barnes Ice Cap (pl. 1, *B*), dowels were set in and measured, but motion in a few days was less than the errors of the method.[8] In many places visited, the upper ice layers were seen to jut out 1 or 2 inches over layers immediately below. Some shear cracks were clearly older, having been transected and displaced by younger and often steeper fractures. In some of the dirty fractures tabular pebbles and boulders showed a preferred orientation parallel to the plane of shear, suggesting differential motion. Not all layers exhibited ice stands out sharply after the winter snow disappears (pl. 2, *A*). The ice cliffs in six lakes show dirt-filled ice at the very lake shores. Above the dirty ice rim, one may travel for tens of miles without finding a sand grain—even less a boulder. There are many clean shear planes in this upper ice, but they are farther apart and less regular in strike and dip as one gets a quarter-mile or more back from the margin. Slightly open steep fractures were encountered, to 12 feet in depth, in an excavation at Ice Camp A2 (fig. 1). The belt of black ice around the rim rises and falls from 1,300 to 2,000 feet in altitude with the marginal topography, but it rarely rises

[8] Most dowels fell out or were broken within a week; motions were insignificant.

more than 200 feet vertically above the flat land around the icecap. In a few places it seems to reflect topography buried just back under the ice edge, for it arches up perceptibly where no hill is exposed. In a few shallow depressions seen in flight, single spots of dirt outcrop as much as a mile back from the ice edge, and from each one a narrow fan of dirt-covered ice reaches to the edge. This suggests a nunatak just submerged.

Because the shear planes slope 10°–36° back under the ice and dirt outcrops generally no more than 200 feet above the ice toe, it stands to reason that the raising of the dirt is from basal ice not very far back under the glacier—theoretically, no more than 1,600 feet from the edge (assuming that the dip of shear planes seen at the surface continues at depth). Turbulence of plastic motions may raise some dirt into the lower few feet of the ice across the whole base of the glacier, because erratics identified as having formerly traveled up to 15 miles toward the northeastern coast were found; but it seems that concentrations of dirt along shear planes do not take place underneath more than 280 feet of ice. This is the depth of ice 1,600 feet back from the toe.[9] Commonly, dirt does not crop out half so high as 200 feet, nor do shear cracks slope so gently as 10°; so the depth of common shearing is more nearly 140 feet. No wonder that most of the rock material raised is of the same lithology as that of outcrops at the ice margin!

COLLECTION OF DIRT DOWN THE ICE SLOPE

The black-ice slope is covered with a film of dirt which is moving toward the ice toe (pl. 2, *A*). This smear of dirt does not average $\frac{1}{16}$ inch thick; an easy blow of the ice ax brings white ice. On hot days (50°–60° F.) the silt and sand could be seen fairly oozing and creeping in small rills. Black-ice areas had an average incline of 15° from horizontal; variations from 6° to 20° were recorded (fig. 3). Fine clay moves off rapidly into regular small streams, spaced 50 feet apart or less. Stones more than 6 inches thick do not transmit much sun's heat through to the ice; hence, while surrounding ice melts down a foot or two, each larger stone stands on a shaded pedestal. Between the end of June, when snow cover melted, and July 19, rock tables 3–15 inches high developed (pl. 2, *B*). Sooner or later each pedestal melts thin, and each stone becomes unseated. It slides off downhill toward the ice margin. Cracks or hollows on the black ice catch the sliding dirt and accumulate a few inches of half-washed silt, sand, and stones. This also "protects" the ice beneath by reducing melting, and after a warm season it becomes a dirt cone or ridge or chain of cones left 20–120 inches high, while surrounding ice melted down. The sides of these cones melt back with uniform slopes at 30° to 38°, suggesting regulation of melting by the angle of repose of the slipping and sliding sediments undermined at the top.[10] By all these motions dirt moves generally toward the ice edge.

Ablation of this black-ice slope is the most rapid of all parts of the glacier because the thin film of dirt absorbs and traps far more of the sun's radiant energy than does the white ice above. Dowels set into drilled holes along profile C on July 18, 1950, and remeasured until ablation slowed down (August 14, 1950),

[9] Based upon bottom profiles worked out by C. A. Littlewood, of Dominion Observatory, Canada, from a gravimetric survey. Parts of these are shown in fig. 3, profiles *C*, *E*, and *F*.

[10] Similar in form and development to dirt cones recently described by Swithinbank (1950).

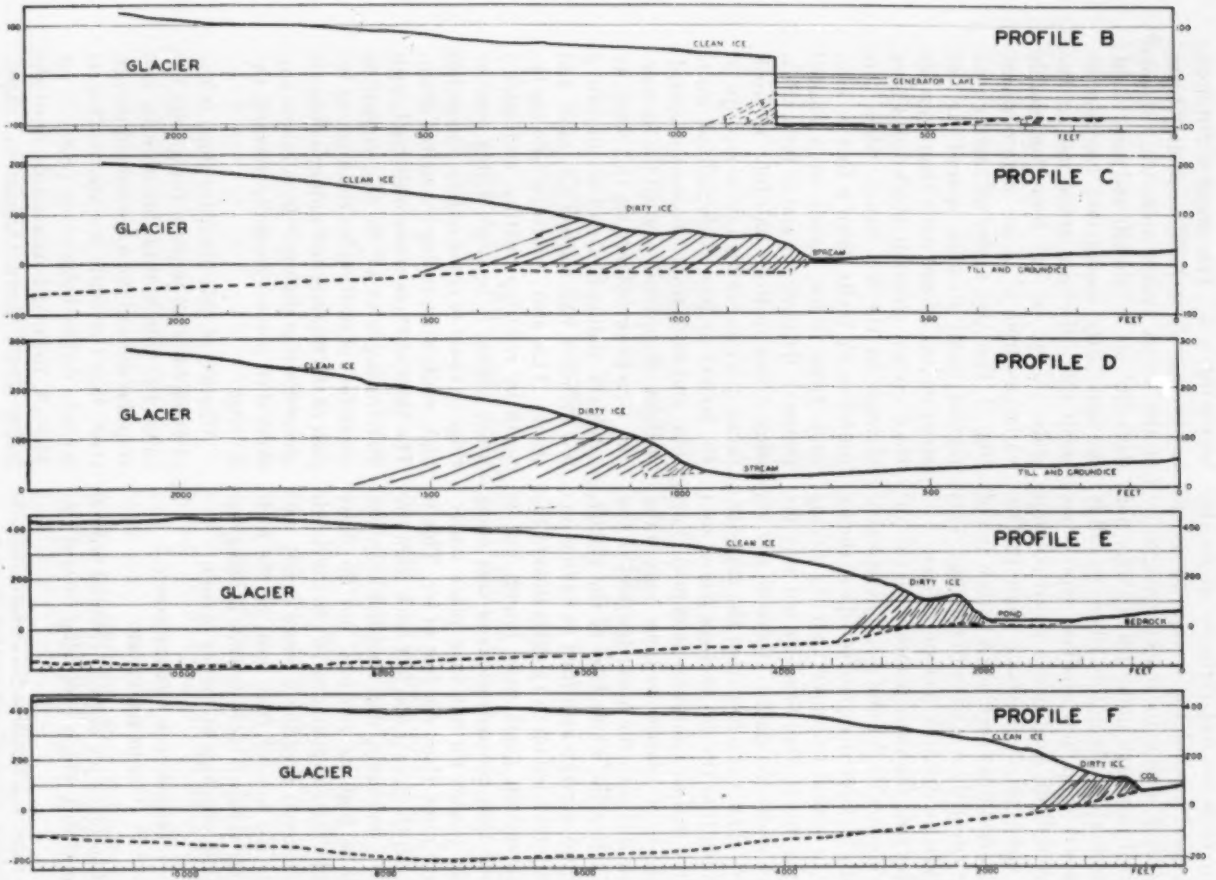

FIG. 3.—Measured profiles across the southern margin of Barnes Ice Cap. Surface determined by hand-level and tape (*B*, *C*, *D*) or by rod and transit (*E*, *F*). Base of the glacier from gravimeter survey by C. A. Littlewood along approximate lines *C*, *E*, and *F*. Base-level = 0 is Generator Lake. Vertical exaggeration 2½ times in *E* and *F* only.

A

B

Stages of marginal moraine formation

melted out at the rate of 8 inches per week on clear ice and 13 inches per week on dirt-filmed ice. As the warmest season between July 1 and August 11 produced the only severe thawing and some snow had to be melted off from the ice surface in this period, the annual loss of black ice is estimated at 60–80 inches, while that of white ice just above is 40–60 inches. This difference in melting rates contributes to the steepness of the ice edge. Tangents to the curved profiles (fig. 3) on black ice were 3° steeper than those on the white-ice slope just above. This change in slope was often a visible sharp break (pl. 2, *A*).

Toward the foot of this steepened black-ice slope the creeping dirt gathers so thick that it protects the ice from annual melting. In some places, where recession may be just beginning or where ice motion just equals ablation, dirt was stacked steeply against the ice toe. Where the recession had gone on further, the ice was coated with 2–4 feet of till for at least 100 feet up onto the steep slope.

DEVELOPMENT OF A TROUGH

Along many miles of the southern and southeastern margins of Barnes Ice Cap there is a depression in the ice at the base of the black-ice slope and parallel to the steep, dirt-covered ice edge. It is frequently 30–90 feet above the flatter ground beyond the ice (fig. 2, "trough"). Between this trough and the outermost steep edge of the glacier there is a "false" ice-cored moraine (pl. 3, *A*). No significant motion was detected in the thick ice of this core. Since the depression, 2–50 feet deep, lies between this ice-cored moraine and the active black-ice slope, it stands to reason that the depression did not exist at first and that there was formerly a continuous convex slope from the black ice to the top of the ice-cored moraine (fig. 2, "previous stage"). Only in this way could the dirt cover of the moraine collect over the lower half of the dirty ice rim for a period of years.

The ice-cored moraine is remarkable, in that the thickness of till cover on the ice rarely varies more than ±1 foot from the average of 3 feet seen at cut after cut and dug or drilled at a dozen localities. This probably reflects the fact that the thaw zone approached this depth early in August and that this is the requisite mantle for effective insulation from summer temperatures. The steep outer face of the moraine, which slopes at 30° (fig. 3) indicates the steepest slope upon which such a protective cover can cling for a few decades. The mechanical composition of the till on profile C has a median near ½ mm. and consists of about 14 per cent silt and clay (below $\frac{1}{16}$ mm.), 20 per cent fine sand ($\frac{1}{16}$–$\frac{1}{4}$ mm.), 34 per cent coarse sand ($\frac{1}{4}$–1 mm.), 15 per cent granules (1–4 mm.), and 17 per cent pebbles (over 4 mm.). Such sandy till with boulders is to be expected in an area of coarse crystalline rocks, especially after ablation seepage has washed out some fines.[11]

[11] Fine sediment brought up from the bottom of Generator Lake, 2,000 feet from the cliff of Barnes

PLATE 3

A, Trough parallel to the ice margin containing a shallow pond. Beyond it is the ice-cored moraine with a breach where the pond has its outlet. Profile *E*.

B, Undermining of the ice core of marginal moraine accumulation. As the dirt-smeared ice cliff melts back, a 3-foot layer of till on top avalanches down the melting slope. At the base, the end moraine is being emplaced.

Why does this trough develop in the black ice? Of the various processes observed, the contrast in ablation rate of the ice-cored moraine, which melted down hardly at all, and the black-ice slope, which melted excessively, is judged to be the major factor. Provided that the annual motion of active black ice out over the toe of dirt-covered ice is not so rapid as the annual ablation of the black ice (60–80 inches per year), the black-ice slope will recede. Thus it seems that, in order to initiate the ice-cored moraine—which is to become a belt of end moraine—there is first a rough balance of discharge of dirty ice and ablation on the black-ice slope (fig. 2, "previous stage"). The duration of this determines the volume and extent of the protective dirt cover. The accumulation of a 3-foot-thick cover over the lower half of the former black-ice slope—this is the common thickness—would take on the order of 25 years. Then it must be followed by a decreasing flow of dirty ice or increased ablation, coupled perhaps with increasing height of dirty shear cracks, to allow the black-ice slope to recede. In so doing, it leaves a shelf or trough between black ice and the ice-cored moraine (pl. 3, *A*).

Once initiated, this trough enlarges by several processes. For one thing, the air in the trough seemed warmer on sunny days. When katabatic winds arose, the trough afforded slight protection. To some extent at least, the quieter air in the trough is heated by radiation from dark slopes on both sides.[12] In addition, most smaller radial streams from the vast interior ice surface flow into the trough wherever there is one. Some are ponded in enclosed hollows in the trough, and this slightly warmed water melts some ice beneath. In the trough each stream flows parallel to the ice edge until it reaches a breach in the ice-cored moraine. Absorption of the sun's heat by dirt in the streams and corrasion by the moving dirt both deepen the channel. Thus the trough may deepen vertically.

The gathering of dirt in the trough from the black-ice slope counteracts the deepening processes to some extent. If recession of the black-ice slope is rapid, there are no extensive dirt accumulations in the trough, so that it widens and deepens continually; in one area the trough was enclosed on both sides to a depth of 74 feet and was about 500 feet broad. This produces a backslope on the ice-cored moraine which is inclined 13°, on the average, into the icecap (fig. 2, "present stage"). If recession of the black-ice slope is relatively slow, sufficient dirt collects in the trough to protect it, and a broad irregular shelf of dirt-protected stagnant ice is added to the ice-cored moraine. One example was 3,000 feet broad; since the dirt cover averaged about 3 feet thick, this might require several centuries to form. Wherever the cover of ablation moraine was broad, the topography was irregular, with minor ridges and hollows marking spasms of retreat of the active black ice.

In most sectors the dirt cover in the bottom of the trough is irregular. It gathers by at least three means: (1) early in July, bands of poorly sorted sand came down the black-ice slope as slush avalanches, ending on snowdrifts and later melting down into the trough; (2) better-sorted layers of silt, sand, and fine gravel

Ice Cap, was 16 per cent clay (to 0.002 mm.) and 84 per cent silt (0.002–0.063 mm.). Washing out of such clays resulted in the very sandy material (less than 3.3 per cent silt and clay) sampled from one dirt cone on a mountain ice field and from two end moraines near Clyde Inlet.

[12] Measurements taken by W. H. Ward (1951) "show that screen air temperature over debris-covered ice was sometimes as much as 2° C higher than over clean ice 70 m away."

washed spasmodically into temporary pools 10–300 feet long in the trough; and (3) till slipped and slid into the trough from melting dirty ice on both sides.

After a few seasons of melting, spots of thick dirt concentration stand higher as dirt cones. Most dirt cones observed in the trough area were larger than those far up on the black-ice slope, suggesting that they formed under a thick dirt blanket and lasted several years. Some dirt-covered ice cones are capped with a till mass, others with the bedded sand of a former pool.

EMPLACEMENT OF THE END MORAINE

When and how does this accumulated load on ice become a deposit on solid ground? The high ice-cored moraines with 30° slopes contrast markedly with the subdued ridges of till and boulders on solid ground (fig. 2, "end moraine"). It becomes evident that ice-cored moraines can lead to these puny end moraines when one realizes that the solid mantle material of the ice-cored moraine is less than one-tenth its height. Invariably, steepness was associated with a high ice core exhibiting the inclined shear planes of formerly active glacier ice. Under the subdued ridges of end moraine, 100 feet to 50 miles away from the steep ice toe, there are layers of clear crystalline ground ice, 1–5 feet thick. In fact, such ice, common to all arctic regions, was also seen under ground moraine and fine outwash generally.[13] But this ice does not contribute to the relief of end moraine.

Various processes destroy the core of glacier ice in the ice-cored moraine. Wherever the circumferential streams in the trough had ponded and broken over the ice-cored moraine a deep, sharp gash developed (pl. 3, *A*). These outlet openings occurred at roughly every 1,000 feet along the ice front. On warm days each one delivered a rushing cataract of water to the outer face of the ice-cored moraine and exposed ice walls with slumping till and boulders.

Elsewhere along the steep outer face of the ice-cored moraine, landslides (mudflow and rock avalanche) were effective means of destroying ice. As thawing of the dirt mantle reached 2–3 feet in depth in July, the soggy till became unstable. In a few places the till crept slowly by solifluction; in others it oozed rapidly by mudflow. The core ice exposed on the upper slope then proceeded to melt rapidly (pl. 3, *B*). On warm days it undermined boulders and stones of the till cover above, and these skidded and bounded down the exposed ice every minute or two. It is necessary only to remove the protective dirt cover here and there to undermine the glacier ice buried under the moraine. Heat penetration of the 3-foot till cover is not necessary, although it may occur. In two places excavation through the covering dirt of the ice-cored moraine to the glacier ice suggests that the thaw layer penetrates to the ice itself by the end of July—at least here and there. Thus the ice core is destroyed, sometimes rapidly, but in any case irregularly.

The texture of the dirt in the final moraine is understandably varied. For the most part, it is the same as the sandy till

[13] Described by many authors and summarized by Black (1950). This interpretation does not agree with Ward (1951), who refers to most buried ice excavated at some distance from the Barnes Ice Cap as "glacier ice." In a few places it was clear that glacier ice was or might be trapped beneath ablation moraines for centuries. The steep topographic rise on ablation moraines covering unequivocal glacier ice was in such contrast with the smooth topography covering widely spaced exposures of clear granular ice layers that most buried ice is believed here to be ground ice (permafrost).

of the ice-cored moraine. Samples from old moraines 40 miles from the Barnes Ice Cap were even more coarse: 8 per cent clay and silt, 6 per cent fine sand, 18 per cent coarse sand, 14 per cent granules, and 54 per cent pebbles. In addition, there are many lenses of well-sorted sand. Logically, these are derived by streams washing them into pools in the trough. Such deposits are too thin and nondescript to merit the topographic description of kame. There is an additional source of coarse gravels and sands on the outer fringes of the ice-cored moraines. The streams debouching through each gorge deposit extensive local fans (fig. 2, "alluvial cone").

Big boulders tend to become concentrated in certain irregular areas of the moraine as it "lets down." At the site of each gorge passing through the ice-cored moraine, large boulders slip into the cut which is transverse to the moraine axis. On the exposed slopes of core ice there is rude sorting, for large boulders bound farther out and fine sediments pile up at the high edge of the heap next to the ice. Thus some initial irregularities of boulder concentration in the low end moraines seem explainable.

Most of the ridges of end moraine near Barnes Ice Cap were no more than 5–15 feet high and 50–500 feet broad. They were first to appear above the snow cover as it melted in June. These ridges occur in nesting groups, but rarely could any one undulating ridge be traced more than a mile across the broad hills. It is to be noted that the distinction between these end-moraine ridges and the intervening ground moraine is only one of relief and magnitude of the local roughness. End moraines representing one simultaneous position of the ice edge stand higher merely because of greater volume. In content, certainly, the upper till of the ground moraine is the same as the till in the end moraine.

The probability that the lower till is plastered onto the rock from beneath the glacier and that it is found under both ground and end moraines is great, but it was not confirmed

Many low end-moraine ridges did not harmonize with the orientation of the present ice-cored moraine. These were covered with well-established tundra vegetation as well as century-old patterned soils, so they are considered very old. However, one area of moraines parallel to the present ice edge along the southwestern side shows three definite belts of recent moraine (pl. 1, *A*): (1) An outer sharp, sandy, boulder-strewn compound ridge from 10 to 50 feet above adjacent older drift and covered with lichen and many plants. No ice was discovered by digging, and there was no sign of recent slump or of steep, ice-cored slope. (2) An intermediate lower belt, dotted by a few very high humps and covered with some lichen. May contain ice locally because some slopes are steep (?). (3) An inner hummocky high belt, continuous and ice-cored, as exhibited in stream gorges. Very few small lichens were present. Dwarf willow was absent. Here perhaps is the "typical" recessional moraine series formed by accelerating and decelerating retreat (fig. 2, "later stage"). In another area on the eastern side of Barnes Ice Cap three belts of moraine, roughly 300 feet apart, all showed buried ice at places. Both by geological estimate in the field and by an independent estimate of the lichenologist, the relative ages of these moraines are twenty-five, ten, and three years, respectively. The chances are that they are double or treble this and may not be fully "let down" to final resting places for one or more centuries still.

CONCLUSIONS

Assuming that the foregoing observations have been synthesized into a proper relationship, these deductions may be made:

1. The material of the end moraines is nearly all local, being derived chiefly from bedrock only a few hundred feet back under the ice from the glacier margin. The few far-traveled erratics must have moved most of the way near the base of the ice.

2. Most of the material is raised to the ice surface along shear planes, which rise toward the glacier margin and seem to penetrate commonly no more than 250 feet below the ice surface.

3. Nearly all the material of end moraines becomes superglacial load (ablation moraine) prior to its deposition.

4. Some of the clay and silt is removed from the ablation moraine by meltwater seepage over the surface on warm days. The till must necessarily become more sandy and stony.

5. Large volumes of end moraine entail many decades or centuries of accumulation, since dirt concentration in the ice is very lean and much ablation must occur.

6. Irregular hummocky final deposition is implicit in the sloughing-off and avalanching down the sides of an enduring ice core.

7. Rude sorting into surficial boulder-rich zones is accomplished wherever the material is emplaced by avalanching down steep faces of the ice core or in meltwater notches through the ice core.

8. Lenses of sorted sand and gravel may be collected in catchment basins in the trough of ice, chiefly on the side of the moraine toward the main icecap. Fans of dipping gravels are often banked up on the opposite outer side of the moraine.

9. The end moraine and the upper till of the ground moraine differ only in the thickness of concentration. Dividing lines between the two are arbitrary and signify principally decelerating or accelerating recession of the ice edge.

10. These sequences and principles apply only to an "active" ice edge in which some motion takes place to within 100 or 200 feet of the ice toe. No wide area of stagnant ice was studied.

ACKNOWLEDGMENTS —Thanks are due to William H. Ward, of England, and John Waller, of Canada, for helpful comments, criticism, and discussion in the field. Much detailed information was gathered by these men at the ice-edge location of Camp A2. The meticulous planning and patient leadership of P. D. Baird made this work possible. Grants-in-aid of this study were made from funds granted by the Ohio State University Research Foundation to the university, for aid in fundamental research, and from the Penrose Fund of the Geological Society of America.

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OGIVES OF THE EAST TWIN GLACIER, ALASKA THEIR NATURE AND ORIGIN[1]

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ABSTRACT

The perfection, fine exposure, and unusual accessibility of ogives on East Twin Glacier, Alaska, make this an exceptional place in which to study the nature and origin of this phenomenon. The origin of ogives is discussed, and previous hypotheses are critically analyzed. With the exception of a hypothesis first suggested by R. T. Chamberlin, that ogives are the surficial expressions of shearing planes, no other hypothesis satisfactorily accounts for the fact that these ogives are exposed edges of layers of denser and dirtier ice than the intervening layers. Concepts of glacier flow evolved by Demorest provide a reasonable mechanism for understanding their formation, namely, periodic obstructed extrusion flow down-glacier from an icefall. Debris which was originally basal is believed to become, by upthrusting and ablation, the surface manifestation of an ogive.

INTRODUCTION

TERMINOLOGY

Agassiz (1847, p. 216)[2] first applied the term "ogives" to the series of curved layers of debris-rich ice whose surface manifestation in the ablation area of a glacier resembles large-scale pointed arches. This usage is believed to have special merit and will be followed in this paper.

Ribboned structure, alternating thin blue and white ice layers, was considered to be a type of ogive by H. and A. Schlagintweit (1850, pp. 78–86, 101), but this broad usage has not been followed and is undesirable. Ribboned structure is a distinct phenomenon in occurrence, if not in origin. Most glaciers form ribboned structure; few form ogives.

The terms "dirt bands" (Forbes, 1842, p. 89), "Forbes' dirt bands" (Quincke, 1905, p. 543; Fisher, 1942, pp. 1–17), and "periodic annual banding" (Washburn, 1935, p. 1885) have been applied to the features herein called "ogives." "Dirt bands" is an unsatisfactory term, because it has also been widely applied to annual dirt layers in stratified firn and ice. These are the German *Schmutzbänder* of present usage. "Ogives" has priority over all the other terms mentioned and has the further advantage of identical form in English, French, and German.

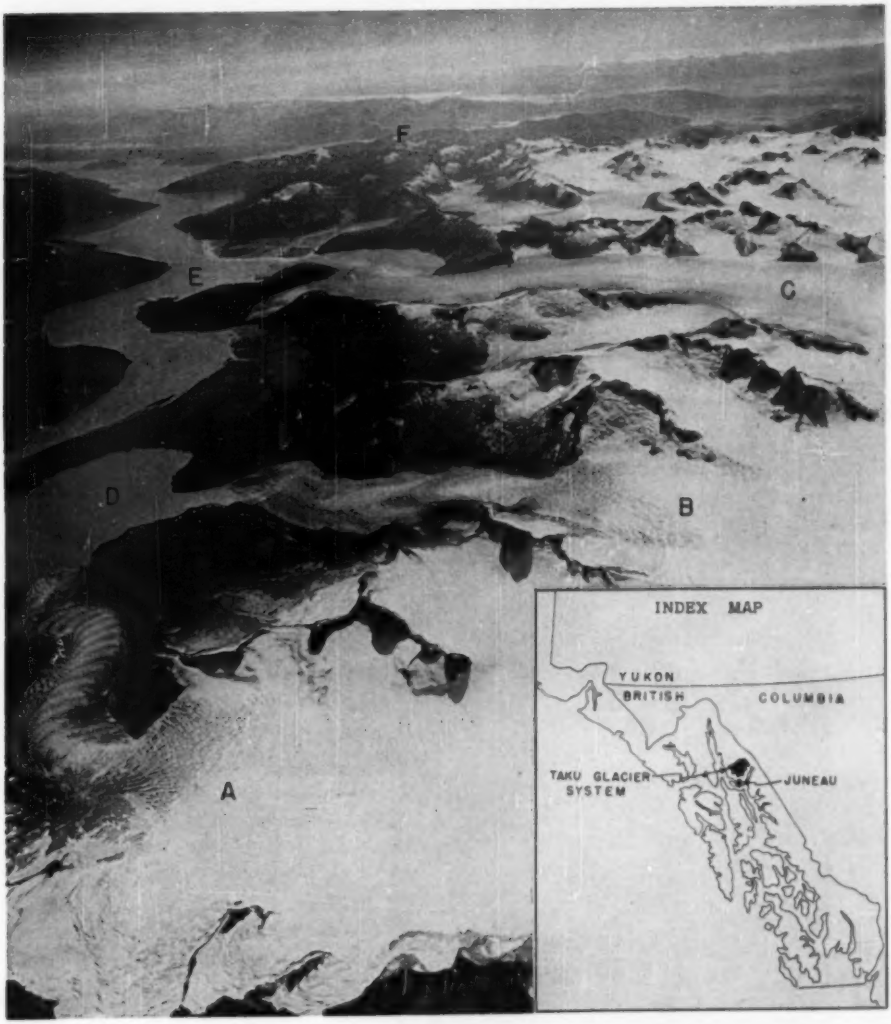
In addition, "periodic annual banding" and "Forbes' dirt bands" are not entirely apropos as terms. The annual formation of an ogive has not been clearly demonstrated, and it is controversial whether J. D. Forbes, for whom "Forbes' dirt bands" were named, was their first discoverer. With the exception of Tyndall (1876, p. 130), nineteenth-century glaciologists attributed the first

[1] Manuscript received February 28, 1951.

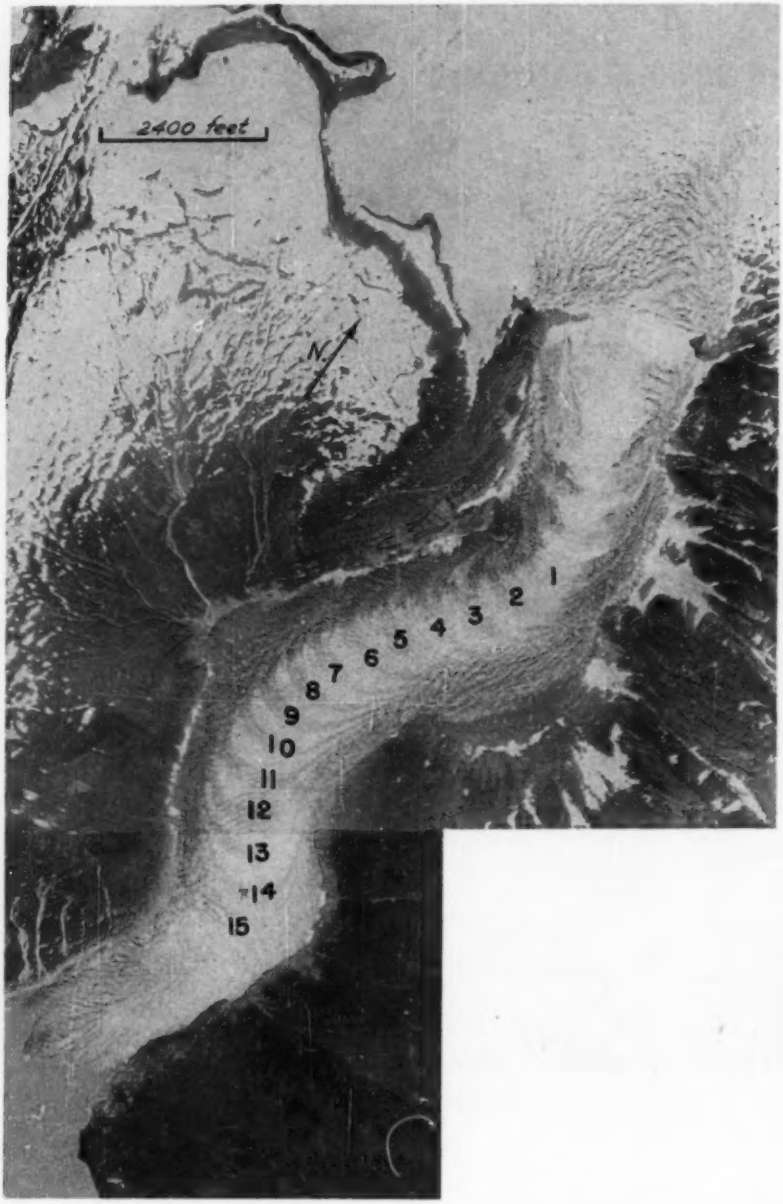
[2] Actually, the term appeared earlier in an article by M. E. Desor (1843, p. 308).

PLATE 1

Aerial view of Taku glacier system, looking southwest. Lettered locations: *A*, East Twin Glacier; *B*, West Twin Glacier; *C*, Taku Glacier; *D*, Twin Lake; *E*, Taku Inlet; *F*, approximate location of Juneau, 24 air miles from Twin Lake. U.S. Air Force photograph.



Aerial view of Taku glacier system



Ogives of the East Twin Glacier

recorded mention of this feature to Guyot (unpublished paper, 1838) and Agassiz (1840, pp. 120–121). The fact that Agassiz and Guyot failed to discriminate between ribboned structure, primary stratification, moraines, and ogives may have been largely responsible for the resulting variance in nomenclature.

Subsequent authors have also confounded ogives with dissimilar phenomena. Russell (1897, pp. 43–44) apparently confused ogives with the outcrops of primary stratification on Sierran glaciers. Fisher (1942, pp. 1–17; 1947, pp. 137–145) confused what he termed "Alaskan bands" with what in reality are ogives.

"Margins" as used in this paper refers to the lateral margins of a glacier. "Lateral moraine" refers to the accumulation of debris on and within a glacier along its margins.

SCOPE OF WORK

This study of ogives is based upon brief observations made during the Juneau Ice Field Research Project of 1949, sponsored by the American Geographical Society and carried out under contract with the Office of Naval Research, with the active co-operation of the Department of the Army and the Air Force, the Forest Service, and other civilian agencies.

Ten days, September 4–13, 1949, were spent on the Twin Glaciers, six of which were devoted to studying ogives. Since then an extensive study has been made of the literature. Although the paucity of field data is recognized, it is felt that the field data and information compiled from published observations as interpreted here in the light of modern concepts of glacial mechanics constitute a contribution to the understanding of ogives.

PHYSICAL SETTING

The East Twin is a small valley glacier, roughly 3 miles long and an average of 1 mile wide. Its source is the same accumulation area as the West Twin Glacier, which is roughly parallel in position and symmetry (pl. 1). Both descend icefalls and calve icebergs into Twin Glacier Lake. These are the only two glaciers in the Taku system that form ogives. Because ogives on the East Twin are regularly spaced and conspicuous and those on the West Twin are irregular and indistinct, the former were accorded particular attention and will be described here.

Roughly 2 miles above its terminus, East Twin Glacier descends 1,100 feet in a horizontal distance of 1,000 feet as an icefall over a steep, narrow cliff. The ogives occur below this icefall. Inasmuch as the zone of the icefall is believed to be a critical factor in their formation, attention will be given it first.

DESCRIPTION

THE ICEFALL AND ITS RELATIONSHIPS

Above the icefall, the East Twin Glacier is broken by tension forces into a great number of curved crevasses pointing up-glacier (pl. 2). No glacier-wide crevasses are present above the icefall; in fact, none extends even half the width of the glacier. In the course of its descent the ice becomes so badly shattered and disrupted that no crevasse retains its identity (pl. 3, *A*). Even seracs in the upper icefall are pulverized by the cascading.

PLATE 2

Ogives of the East Twin Glacier. The definite ones are numbered. U.S. Navy photograph.

Although the firn limit lay at least 1,000 feet above the icefall, the disintegration within the icefall was so complete that all evidence of the dirty summer surface on the glacier was destroyed. Moreover, there was no visible contamination of ice within the icefall by morainic debris to the depth of observation of roughly 75 feet.

In an icefall the ice is necessarily thinner than at places either upstream or downstream therefrom, because increased velocity is accompanied by decreased thickness. Owing to extensive exposures of the bedrock riser which restricts the icefall to a narrow gap, it is not likely that the icefall is very thick. This would imply that the ice is also broken up in depth and that plastic flow, if present at the base, is not considerable.

For 3,000 feet below the foot of the icefall, crevasses are lacking except along the glacier margins. Ogive banding is also absent. The glacier surface is firm, unbroken, and relatively clean, only the margins being strewn with detritus. Roughly 1,500 feet below the icefall the glacier surface becomes mildly undulating. A series of indistinct pressure ridges are faintly outlined by narrow discontinuous white bands, which represent residual accumulations of snow on the south sides of the ridges. Small ponds occupy the depressions between several ridges.

The ridges are 200–500 feet in width and four in number. They are crudely concentric, pointing downstream, and gradually become cuspate toward the glacier sides. Short crevasses cross the crest of the downstream ridge at right angles to its trend.

The ridges die out 3,500 feet below the foot of the icefall, and they are succeeded by the first faint appearance of an ogive. This transition from pressure ridge to ogive is gradual, and the general relations suggest that the down-glacier sides of the ridges become ogives and that the upstream sides, including the narrow intervening troughs, become the cleaner bands. The termination of pressure ridges also marks the beginning of intense transverse crevassing. This is believed to be in response to topographic control, inasmuch as a resistant bedrock ridge projects transversely from both sides of the glacier 3,500 feet below the icefall and narrows the glacier perceptibly. Beyond this ridge the glacier floor again appears to increase in gradient.

OGIVES

East Twin ogives are most striking when viewed from an elevated position (pls. 1 and 2). Traced downstream, they at first become increasingly distinct, but farther down they gradually become less distinct. Fifteen successive ogives were definitely identified and are numbered in plate 2. Pressure ridges just below the icefall are believed to be incipient ogives. These and the remains of several other ogives in the terminal portion, if included, would bring the count to twenty ogives in roughly 2 miles of glacier.

PLATE 3

A, Middle portion of East Twin icefall.

B, Lower East Twin Glacier viewed from west bank. Discordance between bands at far right is an optical illusion caused by foreshortening and unevenness in the glacier surface.

C, Longitudinal section of East Twin terminus. Remnant ogive stratigraphically below most prominent debris layer.

D, Ogive ice on left; bubbly, white ice on right. Twin Glacier.

B

D

A

C

Icefall and ogives of the East Twin Glacier

Plate 2 reveals the remarkable regularity of size, shape, and spacing of the ogives. They are mostly 250–350 feet wide in the center of the glacier. This approaches the average width of the intervening areas. Ogives near the icefall are lunate, becoming gradually more parabolic toward the terminus. Unlike the ogives viewed from air photos of other Alaskan glaciers, these did not become more closely spaced or more hyperbolic toward the terminus. The ogives die out before the terminus is reached, as a result of intense crevassing preparatory to iceberg calving.

All narrow toward the glacier margins, and most of them lose their identity upon merging with the morainic debris of the margins. On the east side of the glacier, where the margin is relatively clean, ogives *8–11* could be traced up-glacier as narrow subparallel bands, which finally coalesced. It was possible to follow ogive *11* to a point opposite the apex of ogive *5*, a distance of almost 3,000 feet.

Lateral morainic detritus is much more prominent on the west than on the east margin of the glacier (pl. 3, *B*). Aprons of avalanche debris on the west margin at the base of the icefall have been dragged into arcuate stringers by the more rapid movement of the center of the glacier. Farther down-glacier, concentrations of morainic detritus emphasize and coincide with the west sides of the ogives. It is also present in intervening areas, but in most cases it does not extend so far toward the center of the glacier in these areas (pl. 3, *B*). Hence the ogives appear to incorporate more morainic detritus than do the intervening areas.

A traverse was made of ogives from the terminus of East Twin Glacier to ogive *8* (pl. 2). Measurements and particular observations were made of ogives *8*, *9*, and *10* and of the intervening areas. The distinction between ogives and intervening areas is relative only, but the following general differences are noted:

1. Ogives are composed of denser and dirtier ice than the intervening areas (pl. 3, *D*). The difference in color is imparted by the concentration of debris in ogives and by the contrast between the predominantly bubble-free blue ice of ogives and the predominantly bubbly white ice of the intervening areas.

2. Debris within the ogives is more abundant, coarser, and more evenly distributed than in the intermediate areas.

3. Owing to differential ablation caused by a difference in albedo, the cleaner and whiter intervening areas project a few inches to a foot above them (pl. 3, *D*).

Ogives *8*, *9*, and *10* are nearly spoon-shaped structurally, dipping 70°–80° toward the center of the glacier at the margins and 35°–50° up-glacier at the vertices of the ogives. They are conformable with the intervening layers and with the ribboned structure, which is less conspicuous in ogives than between them. The average true thickness of ogives *8*, *9*, and *10* measured 240 feet and that of the intervening areas 255 feet.

On a small scale, ribboned structures resemble ogives, inasmuch as dense, bubble-free, blue-ice layers alternate with bubbly white-ice layers. The blue layers of ribboned structure have long been attributed to differential movement accompanied by recrystallization through refreezing of meltwater produced by friction along shear planes (Philipp, 1920, p. 439). For this reason the blue ice of ogives is also considered ice that has recrystallized in this manner, with elimination of air bubbles.

Excavations showed that the surficial debris of ogives is derived from the ice by ablation and not from an extra-glacial

source. The surficial debris is predominantly angular and poorly sorted. Sizes range from clay and silt to boulders several feet in diameter. The clay and silt are rather uniformly distributed, and the coarser debris is localized along certain layers. Concentrated superglacial deposits of coarse sand and pebbles form debris cones. Striated cobbles and boulders suggest that some of the detritus was once subglacial. Ogives also contain bubbly-ice layers up to 25 feet thick which are mostly debris-free, but a few contain dust and silt. By contrast, all the blue ice contains layers of debris parallel to the ogive structure.

TABLE 1

SECTION OF PART OF AN INTEROGIVE ICE LAYER

Type of Ice	True Thickness (Feet)
Clean, bubbly white ice	12
Bubbly white ice with many parallel thin dirt layers	4
Dirty blue-ice lens	1
Clean, bubbly white ice	14
Bubbly white ice with many parallel thin dirt layers	4
Dirty blue ice	2
Clean, bubbly white ice	18
Total	55

Dirty ice in the relatively clean ice areas between ogives is of two types: first, parallel dust and silt layers 0.25–0.50 inch thick within white, bubbly-ice layers and, second, debris-laden blue-ice layers and lenses. The debris in nonogive areas is similar to that in ogives, but less abundant. The clean layer between ogives *9* and *10* measured 306 feet in width and 252 feet in true thickness. The section in table 1 of a part of this layer adjacent to ogive *9* is believed to be representative.

The dirt in all the glacier ice is so small in amount that it has often been mistakenly assumed that the surficial debris must have been introduced onto the glacier surface from an outside source instead of being concentrated by ablation (Forbes, 1842, pp. 348–352; Tyndall, 1876, pp. 127–132; Scherzer, 1907, pp. 50–60). The difficulty in tracing the ogive and nonogive dust and silt layers in depth is analogous to the difficulty in detecting annual dirt layers in pits dug in firn. In such pits it may be several days before ablation concentrates the debris sufficiently so it is visible on the pit walls.

Where East Twin Glacier enters Twin Glacier Lake, a striking longitudinal section of the glacier is exposed (pl. 3, *C*). Debris planes are clearly shown, dipping upstream. The most prominent debris plane pictured is believed to be the upper surface of an ogive, because, stratigraphically, below this debris plane is the dense, dirty, schistose ice characteristic of ogives and above the debris plane are the bubbly, relatively debris-free characteristics of ice between ogives.

HYPOTHESES OF ORIGIN

Ogives have been reported from valley glaciers in the Alps, the Himalayas, Canada, the United States, Alaska, and Patagonia and are therefore a feature of wide geographical distribution. Six contrasting hypotheses have been advanced to explain the origin of ogives, as follows:

1. Ogives are original dirty layers in the firn carried through an icefall or restricted channelway and eventually exposed by ablation in the lower part of the glacier.

2. Ogives are surficial bands related to alternating zones of porous and compact ice formed at the base of an icefall, with dirt of surficial origin lodging in the porous ice more readily than in compact ice.

3. Ogives are formed by surficial concentration of debris in depressions at and below an icefall.

4. Ogives are glacier-wide blocks of dirty ice which are separated by the clean, snow-filled crevasses of an icefall.

5. Ogives are formed in an icefall by alternate winter snowfalls and summer dust accumulations.

6. Ogives are the surface expressions of shearing planes and associated phenomena of glacial flow.

All hypotheses were first developed from study of ogives in the Alps, with the exception of no. 5, which has been applied only to the ogives of East Twin Glacier, Alaska. Hypotheses 3 and 4 have subsequently been applied to Alaskan ogives. All except hypotheses 2 and 6 have had proponents within the last two decades.[1]

ORIGIN OF OGIVES OF EAST TWIN GLACIER

Any plausible explanation of the East Twin ogives must account for the following characteristics: (1) location down-valley from an icefall in the ablation area; (2) continuity across the width of the glacier; (3) three-dimensional character; (4) regularity in size, shape, and spacing; (5) thickness in excess of 200 feet; (6) dip up-glacier and inward from the margins; (7) complete disruption of pre-existing stratification in the icefall; (8) gradation from pressure ridge to ogive; (9) layers of dense, bubble-free, seemingly recrystallized ice; (10) coarseness of detritus; (11) abraded clastics; (12) increasing prominence and increasing abundance of detritus down-glacier to the terminus.

1. *Stratification hypothesis.*—This is the only hypothesis which does not demand the presence of an icefall up-glacier from the ogives. This hypothesis, originally proposed by Agassiz (1847, pp. 205–218), was subsequently supported by Hess (1904, pp. 169–178) and Vareschi (Godwin, 1949, pp. 325–332).

Hess formed miniature ogives experimentally with layers of red and white wax. From these experiments he concluded that parallel horizontal firn strata, moving from a wide firn area into a constricted valley, would become thickened and spoon-shaped and that the dirty layers would be accented through the removal by intensified ablation of the surficial seracs produced in an icefall.

Vareschi's pollen studies on the Aletsch Glacier, Switzerland, lend support to the stratification hypothesis. Samples of pollen collected from successive dark and light bands are interpreted as showing a seasonal sequence, the dirty ice representing summer, spring, and autumn accumulations and the cleaner ice corresponding to winter accumulations.

The complete disruption of pre-existing stratification by East Twin icefall definitely eliminates this hypothesis from consideration here. Other facts just as fatal to the hypothesis are (1) the 200-foot-plus thickness of some ogives; (2) the coarseness of detritus; and (3) the failure of the theory to recognize the dynamics of glacial flow.

2. *Porosity hypothesis.*—J. D. Forbes (1842, pp. 348–352; 1859, pp. 17–26, 210–220, 268; 1900, pp. 154–162, 379–380) was the lone proponent of this hypothesis. He was the first to recognize an icefall as a general requisite to the formation of ogives.

The presence of alternating zones of porous and compact ice on the East Twin was verified, but in the reverse relationship from that proposed by Forbes. The ogives are the compact, not the porous, zones. In addition, the porous

[1] W. V. Lewis (in Orowan, 1949, p. 230) by inference seems to favor a shear origin.

zones are not perforated enough to collect particles of rock, and the compact zones are not smooth enough to be washed clean by rain, as Forbes visualizes.

Other facts which oppose Forbes's hypothesis are (1) the East Twin ogives have a three-dimensional character; (2) the surficial debris increases in abundance down-glacier; (3) the amount of debris bearing the characteristics of subglacial action is greater than would be expected to be introduced in the icefall zone.

3. *Depression hypothesis.*—Tyndall (1876, pp. 127–132; 1896, pp. 372–375) believed that Forbes had confused cause and effect and that the greater porosity observed in the ogives was merely the result of the ice surface's being pitted by dirt. His explanation involves three considerations: "(1) the transverse breaking of the glacier on the cascade, and the gradual accumulation of the dirt in the hollows between the ridges; (2) the subsequent toning down of the ridges to gentle protuberances which sweep across the glacier; (3) and the collection of the dirt upon the slopes and at the bases of these protuberances."

Washburn (1935, pp. 1885–1889) followed this thesis in its entirety, apparently failing to recognize that Tyndall had introduced the idea years before. Though Tyndall merely speculated on the annularity of ogives, Washburn was convinced of it.

Scherzer (1907, pp. 50–60, 118–119), Tutton (1927, pp. 113, 121–22), and Streiff-Becker (1943, pp. 111–132) favored the accumulation of debris in troughs at the base of an icefall. Scherzer held that the troughs were due to the incomplete healing of transverse crevasses in an icefall. On the other hand, Streiff-Becker considered that the undulations at the base of the icefall were formed from strong differential pressure from above (Fisher, 1947, p. 140) and by the damming-up of the glacier below an icefall through a narrowing of the glacier bed (Seligman, 1949, p. 330). Tutton (1927, p. 121) regarded the swells as "due to the forces at work in effecting the stream-like movement of the glacier, faster in the middle and at the surface than at the sides and bottom (bed), rather than to mere change of declivity of bed."

Application of this hypothesis to the East Twin ogives is weakened by the lack of glacier-wide crevasses and block ridges in the East Twin icefall and the lack of surface debris available at the icefall. Troughs exist below the icefall, but there was no sign of dirt collecting in them. On the contrary, the pressure ridges appear darker than the depressions.

The increasing abundance of debris down-glacier is also a cogent argument against the accumulation of debris in troughs. Had the debris collected in icefall crevasses, progressive ablation down-glacier might well remove a sufficient thickness of ice to bring most or all of the debris to the surface by the time it reached the glacier terminus; yet all ogives develop and retain a well-defined three-dimensional form despite unlimited ablation.

4. *Block-ridge hypothesis.*—Fisher (1947, pp. 137–145), in contrast to Washburn, believed that the block ridges become ogives and that the glacier-wide crevasses become white bands. In his opinion (1947, p. 140), "the dark bands are the lineal descendants of the old (dirty) névé surface above the icefall, and the clean bands are inlays of clean ice that filled up the crevasses."

The facts which prevent acceptance of

this theory include: (1) lack of glacier-wide crevasses and dirty block ridges in the icefall; (2) coarseness of the debris in the ogives; and (3) increasing clarity of ogives down-glacier. In addition, the "inlays of clean ice" in 75-foot crevasses would be depleted by ablation before the bands reached the glacier terminus, assuming an arbitrary velocity of flow of 300 feet per year and an ablation rate of 2 feet per year.

5. *Snowfall-dust accumulation hypothesis.*—After observations of ogives on the East Twin Glacier, Miller (1949, p. 25) reports:

> It seems fairly clear that the alternation of bands represents, in turn, the white clean ice of winter snow carried over the seracs from late autumn to spring, and darker bands of ice made dirty by accumulations of summer dust carried through the seracs in that segment of the glacier which moved over this point during the warmer months. This summer dirt is concentrated by virtue of ablation and a narrower channel of flow below the icefall.

This concept ignores the fact that much of the debris is not dust, that it is much greater in quantity than any seasonal accumulation, and that much of the debris shows subglacial abrasion. Furthermore, this hypothesis assumes that the area of dust accumulation would be limited to that portion of the glacier's surface which had not moved "over the seracs from late autumn to spring." It also implies that the glacier descends the icefall in roughly half a year. If this premise were true, each ogive should be over 500 feet wide. In all fairness to Miller, he also suggested:

> It is considered very desirable for next season's field party to investigate closely the possibility that here, shearing at depth, may be a factor in bringing out the pattern of dirty summer ice, intercalated with cleaner winter ice, resulting from possible thrusting of many thin layers of ice one against another.

6. *Shearing hypothesis.*—Although T. C. Chamberlin (1895, pp. 206–208) did not discuss ogives as such, his explanation of the means of introducing debris into the small-scale foliation within the ice of ablation areas is especially significant:

> The original stratification could not have been pronounced. Perhaps it was intensified somewhat during subsequent consolidation, but some new agency was necessary to produce the more definite partings and to introduce the layers of debris. This agency appears to have been a shearing movement between the layers.

According to R. T. Chamberlin (1928, pp. 24–26), "where an icefall behind develops strong thrusting, relief is obtained by upward movement of the ice mass as well as by forward movement. This is both by granular adjustment and by shearing." R. T. Chamberlin did not distinguish between the large-scale ogives and smaller-scale foliation, nor did he attempt to account for the periodicity of ogives. He regarded each ogive as a special surface expression of many closely spaced shearing planes.

This mechanism accounts for features not readily explained by other hypotheses. The three-dimensional character of the ogives, their form, the general coarseness of debris, the similarity to thrust layers exposed in longitudinal section in the terminus of the glacier (pl. 3, *C*), and the similarity of the ice in ogives and in the blue bands of ribboned structure, all attest to a shearing origin. Furthermore, the occurrence of pressure ridges at the base of the icefall is an intimately related phenomenon produced by the dynamics of the glacier.

Concepts of glacier flow proposed by Demorest (1942, pp. 29–66; 1943, pp. 363–400) lend support to the shearing hypothesis. His analysis of glacier flow establishes four types of ice movement,

each characteristic of different conditions of underlying topography and surface configuration. These are extrusion flow, obstructed extrusion flow, obstructed gravity flow, and gravity flow. Extrusion flow is "pressure-controlled" because it depends upon differential pressures arising from the slope of the glacier surface. Gravity flow is "drainage-controlled" because it results from the force of gravity, whose effect is dependent upon the slope of the glacier floor.

East Twin ogives seem best explained by the periodic occurrence of obstructed extrusion flow at the base of an icefall. A condition of obstruction to flow, such as is present here, was suggested in principle by R. T. Chamberlin, namely, the sudden flattening of bed slope beneath an icefall.

In making its steep descent, the ice in a fall is accelerated by gravity and attains a high velocity. This velocity increases with the height and gradient of the icefall and the constriction of the cross section of flow. At the base of the icefall this motion is in part arrested, and the relatively level portion of the glacier has to bear the thrust of the steeper portions behind. This obstruction causes the flowing ice to pile up below the icefall and increase its thickness. Compressive stresses accumulate faster than they can be accommodated by plastic flow during the thickening process. Under these conditions the slower-moving thinner ice ahead is overridden by the obstructed ice. Movement of the thrusted layers is forward and upward, deforming the upper zone of relatively nonplastic ice and forming transverse pressure ridges.

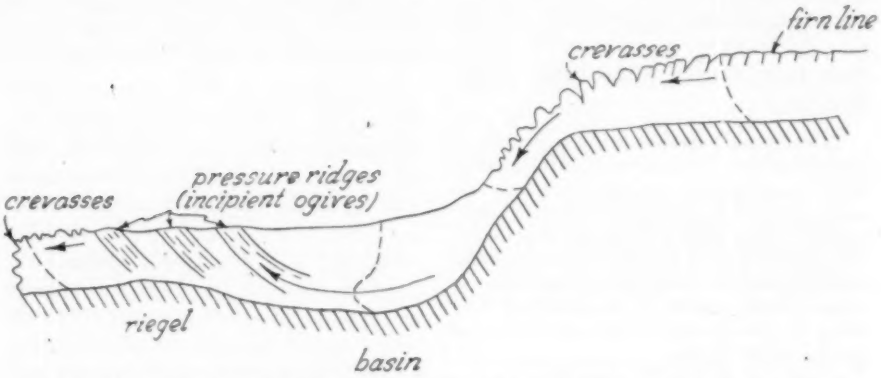


FIG. 1.—Diagrammatic longitudinal section of East Twin Glacier icefall. Dashed lines show profile of rate of flow; arrows show the locus of maximum velocity in different parts of the glacier (after Matthes and Demorest). Ogives form below icefall where flow is obstructed by slower-moving masses.

The absence of tension crevasses from the base of the East Twin icefall to the constriction 3,500 feet down-glacier indicates the absence of gravity flow. Thus the type of flow, we infer, changes from "drainage-controlled" in the icefall to "pressure-controlled" below the icefall, and the level of maximum velocity shifts from the surface to the bottom of the glacier (fig. 1).

It is a matter of common observation that lakes are often located below bedrock steps in valleys of homogeneous rock. Pressure-controlled flow down-glacier from the East Twin icefall implies

that the glacier floor is either basin-shaped or has a gradient too low to enable the force of gravity to overcome the frictional resistance of the rock bed. It seems likely that the former is the case. The bedrock restriction which projects laterally into both sides of the glacier and marks the boundary between the continuous compact ice and the crevassed ice may be a surface indication of a basin sill. If this restriction exists as a riegel (buried ridge) at the lower end of the basin, it may be a factor influencing the resistance to thrusting and hence may have a part in controlling the spacing and regularity of ogives.

In all probability, the shearing and thrusting occur near the lower end of the smooth reach below the icefall, where pressure ridges are most prominent and where the transition from bottom to top current takes place (fig. 1). It is here that what was originally subglacial detritus is believed to become, by upthrusting and ablation, the surface manifestation of an ogive.

PERIODICITY OF FORMATION

Because the ogives on the East Twin Glacier are rather uniformly spaced, a roughly periodic relation must exist between velocity of flow and episodes of thrusting. Presumably, the obstruction below the icefall is a fairly constant force, and the stresses which produce the obstructed flow are accumulative until that type of flow occurs. The stresses are not relieved solely by thrusting of the bottom current, because small-scale shear foliation (ribboned structure) of conformable orientation is also found in the ice layers between ogives. These represent minor yieldings along planes parallel to the ogives and are a product of the same applied stresses. Hence the over-all thrusting is a general periodic movement involving thick masses of ice which carry along thinner slices in the direction of easiest relief, which in this case, owing to the obstruction, is surfaceward. This relieves the obstructing effect temporarily, and plastic flow then dominates. When stresses again accumulate to a point where plastic flow is overcome, thrusting again takes place.

Obviously, it would be wholly fortuitous under the shearing theory if East Twin ogives were formed only by annual thrusting, although annual changes in ice volume or velocity of flow might be an important factor. Glacier flow is sometimes accelerated above firn line during the winter, owing to heavy snowfalls (Streiff-Becker, according to Seligman, 1947, p. 16; Orowan, 1949, pp. 232–233), and it may be retarded at the same time of year below firn line, where accumulative snowfalls are less and where a proportionately thicker surface layer yields less readily in subfreezing weather (R. T. Chamberlin, 1936, p. 104). Such a difference in rate of flow during the winter on East Twin Glacier would increase the obstruction below the icefall and possibly influence the regularity of large-scale thrusting at this point. Seasonal observations, if and when possible, should contribute to this problem.

SCARCITY OF GLACIERS WITH WELL-DEVELOPED OGIVES

Glaciers with well-developed ogives are relatively uncommon as compared to glaciers with indistinct and deranged ogives. From a comparison of the two Twin Glaciers and a study of available photographs of ogives, the following possible explanations for the absence, indistinctness, or derangement of ogives below icefalls are suggested:

1. The glacier is of too great thickness or is too short to permit obstructed ex-

trusion flow in depth to be revealed by ablation.

2. The icefall is composite or complicated by other icefalls, which disrupt the regularity of thrusting at the base.

3. No abrupt flattening of bedrock slope occurs below the icefall.

4. The icefall lacks the size or vigor to develop large periodic thrust movements at its base.

5. At the base of an icefall the glacier makes a sharp turn.

6. A tributary glacier may disrupt or conceal ogives in a trunk glacier.

7. If the distance from an icefall to the glacier terminus is not great, thrust layers formed near the glacier terminus as a result of thinning of the ice may derange thrust layers formed at the base of the icefall.

8. No portion of the glacier below the icefall is in the ablation area.

SUGGESTIONS FOR FURTHER STUDIES

It is natural to refer a phenomenon to a single cause, even though it is rare that any phenomenon is wholly the result of a single cause. If it is demonstrable that this cause is effective in one area, the tendency is to extend its application elsewhere. Therefore, the writer suggests careful study by others of the following points.

For those ogives occurring down-glacier from an icefall the writer is inclined to believe that the factor of thrusting may be universally applicable. That all ogives are related to an icefall has not been established, although this is suspected from studies of the aerial photographs of nine Alaskan glaciers which display ogives below an icefall. Further observations in other glacier regions are needed to prove this.

Studies made at the bottoms of steep declivities in recently deglaciated valleys may aid in reconstructing the flow characteristics and indicate whether topographic and geologic relations would have favored obstructed extrusion flow.

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CRYSTAL FABRIC STUDIES ON EMMONS GLACIER MOUNT RAINIER, WASHINGTON[1]

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ABSTRACT

Optical orientations of 1,725 ice crystals were determined at eleven locations on the Emmons Glacier by means of a 6-inch universal stage mounted between crossed polaroid sheets. The crystals were 0.2–6 inches across and from three to forty were included in each 4½×6-inch thin-section. Down-valley glacier movements of 3.4–9.2 inches per day were recorded at four of the higher locations, and another three were in ice believed to be essentially stagnant.

The optic axes at most locations, when plotted on a Schmidt equal-area projection, form consistent patterns featuring four strong maxima at the corners of diamond-shaped quadrangles. Concentrations as high as 26 per cent in 1 per cent of the area were recorded. The angular separation of opposite corners of the diamond-shaped quadrangles average 84° and 52°, respectively. The pattern in stagnant ice is the same and as strong as the pattern in moving ice. Poles to shear planes and clear ice bands in the glacier at the locations studied fall near the centers of the diamonds. If this pattern is controlled by gliding within the ice crystals, it suggests the possibility of glide planes in ice crystals other than the well-known basal glide plane.

INTRODUCTION

GENERAL STATEMENT

Crystal fabric studies offer a promising method of attack on the still troublesome problem of ice flowage. Glacier ice has recrystallized and flowed under stress and is therefore truly a metamorphic rock. Application of the techniques of structural petrology to studies of ice may eventually contribute to the larger problem of solid flowage in the more durable rocks. Because glaciers are monomineralic and flow on the surface of the earth rather than at great depth, they provide ideally simplified subjects of study in comparison with the more complex metamorphic rocks. Surprisingly little has been done along this line, the work of Perutz (1940) in the Alps being perhaps the best known. The extensive studies of the late Max Demorest regrettably are largely unpublished, and the more recent work of Bader is appearing in this issue of the *Journal*.

The studies reported here are little more than exploratory, but they demonstrate impressive fabrics within glacier ice and raise a number of intriguing questions. The fabric studies were supplemented by observations of surface movement of the glacier at several points.

LOCATION AND PHYSICAL SETTING

Emmons Glacier on the northeastern side of Mount Rainier, Washington (fig. 1), is the largest of the twenty-six glaciers flowing radially down the flanks of this large volcanic cone. It is nearly 6 miles long and flows in a northeasterly direction, confined at lower elevations by steep canyon walls and moraines. Most of the lower 2 miles of the glacier is covered with ablation moraine, consisting of detritus from the finest up to blocks 10 feet or more in diameter. Two tongues of faster-moving bare ice extend down into the debris-covered area along the southeast side (pl. 1, *A*). The terminal and the northwest marginal parts of the glacier have an exceptionally thick cover of debris and are believed to be essentially stagnant. Geological and historical evidence (Matthes, 1944, p. 681) indicates that Emmons Glacier has been retreating

[1] Manuscript received March 2, 1951.

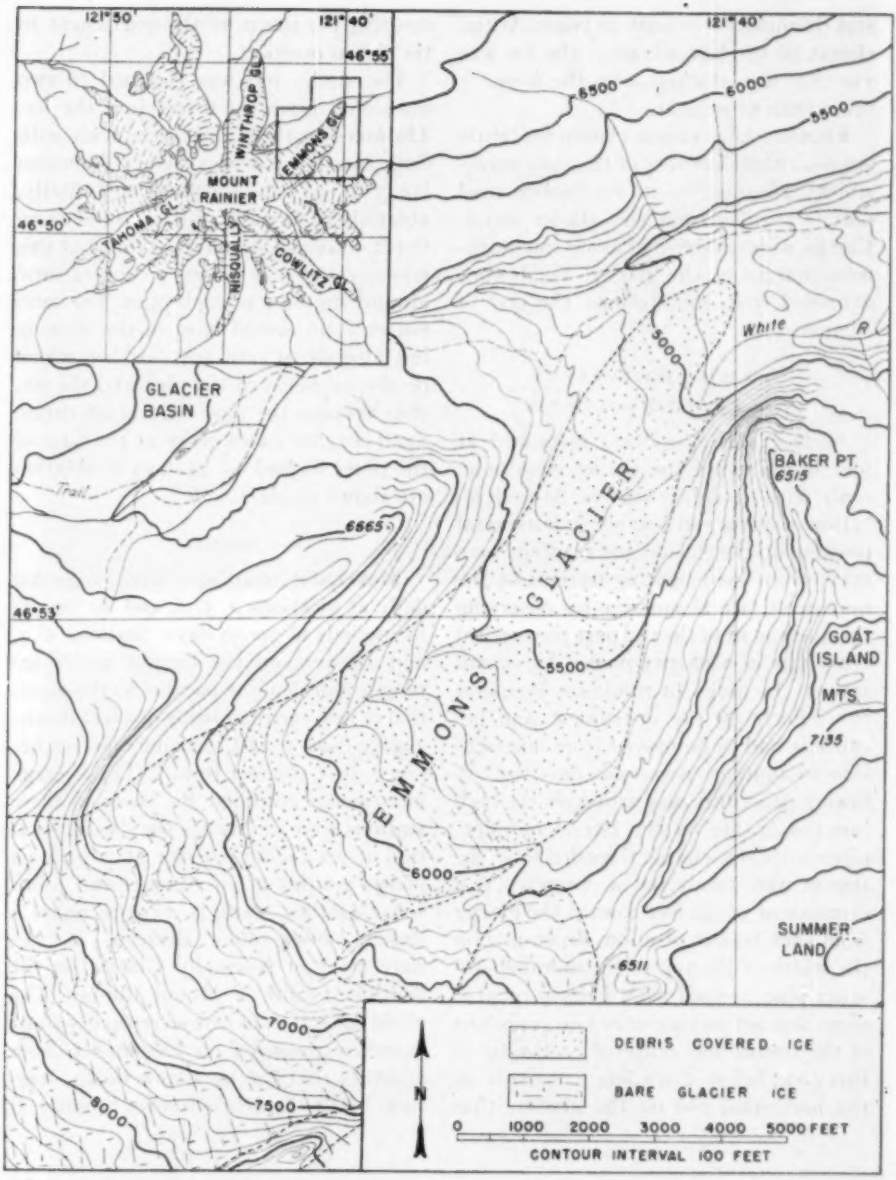


FIG. 1.—Map of Emmons Glacier and vicinity. Modified from U.S. Geological Survey map of Mount Rainier National Park.

and thinning for at least 50 years. At the climax of the last advance, the ice was 150–200 feet thicker over the lower 2 miles than at present.

Emmons Glacier was chosen for fabric studies chiefly because of the easy accessibility afforded by an automobile road only 1.5 miles from the glacier snout. Camps were established along the northwest margin of the glacier. Field work extended from mid-July to the end of August, 1950.

GLACIER MOVEMENT

INSTRUMENTS AND PROCEDURES

Surface velocities were measured at four locations on the glacier simultaneously with the fabric studies. At each location a stadia rod was set in horizontal position on the glacier, and readings were taken from fixed sighting stations on the northwest lateral moraine by observing the movement of the rod past the vertical cross-hair of a twenty-power Ainsworth transit. In order to minimize temperature effects on the instrument and because it had to be moved from one location to another twice each day, the following procedure was arranged. At each instrument site on the lateral moraine, pipes with brass heads threaded to fit the transit were firmly set in concrete. At a distance of 30–50 feet toward the glacier from each transit position, depending on the width of the top of the moraine, another pipe, topped with a sharp-pointed cone, was set in concrete. The cross-hair of the transit was centered on the tip of this cone before a reading was made on the horizontal rod on the glacier, thus assuring the return of the instrument to the initial position.

The stadia rod was clamped to two aluminum pipes set 6 feet into the ice. The holes for these pipes were made with drills based on a design used by Ahlmann (1935, p. 31) in Spitsbergen. Initially, ablation around the pipes allowed them to tilt, making the rods unstable, but this was overcome by piling a cone of sand around the base of each pipe. The sand not only prevented most of the melting but also sifted into any opening which developed between the pipe and the ice, thus keeping the pipe rigid at all times. Sand-covered cones of ice at the base of the pipes formed by 12 days of ablation are shown in plate 1, *B*.

RESULTS

Movement readings were recorded daily at locations *2*, *5*, *8*, and *10* (fig. 2) for periods of 20–30 days. Stations *5*, *2*, and *10* were on the longest active ice tongue and in a line parallel to the direction of movement, *2* being 900 feet down-glacier from *5*, and *10* being 660 feet below *2*, or 1,560 feet from *5*. The average movements recorded for each of these stations were 9.2 inches per day at location *5*, 7.1 inches per day at *2*, and 3.4 inches per day at *10*. Stagner and Johnson (Matthes, 1946, p. 223) recorded a similar down-valley decrease in the movement of Nisqually Glacier on the southwest flank of Mount Rainier. The rapid reduction in rate over such a short distance measured on Emmons Glacier suggests that the ice is practically stagnant below the debris-free ice tongues. It

PLATE 1

A, Lower part of Emmons Glacier viewed from the northeast. Photograph by B. H. Bieler.

B, Stadia rod for movement studies as set up on glacier. Debris-covered cones at base of pipes show method of stabilizing pipe. Distance from surface to man's finger on pipe demonstrates ablation in 12 days.

A

B

Emmons Glacier and equipment for movement studies

A

B

Methods and equipment used in fabric studies

also implies surfaceward movement of the ice (Demorest, 1942, p. 45), for gradual conversion of the horizontal component of flow into an upward direction appears the best mechanism for explaining the down-valley decrease in horizontal movement. A possible structural indication that this actually happens is afforded by clear ice bands, 0.5–1 inch wide, dipping up-glacier at a steep angle.

Daily movements were also recorded at location *8* on a medial moraine northwest of station *2*. The down-valley movement here averaged 3.5 inches per day, confirming the impression previously gained from the ablation mantle that this ice was less active than in the clear ice streams.

ABLATION

Whenever pipes were sunk into the glacier to support stadia rods, marks were made on them level with the ice surface, in order to record ablation. Plate 1, *B*, shows the amount of ablation after 12 days at location *10*. The average daily ablation measured at the three movement stations (locations *5*, *2*, and *10*) on active ice was 3.1 inches of ice.

GLACIER FABRIC

INTRODUCTORY STATEMENT

Ice, at the pressures and temperatures attained in glaciers, is hexagonal, probably in the ditrigonal pyramidal class (Palache, Berman, and Frondel, 1944, p. 494). Thin-sections were made, and the orientations of the optic axes of the crystals were measured at eleven locations on Emmons Glacier by means of standard universal-stage procedures. Six of these—locations *2*, *3*, *4*, *5*, *10*, and *11*—are on the longer of two tongues of clean active ice in the central part of the glacier (fig. 2). Two—locations *8* and *9*—are about 100 feet apart on a medial moraine north of location *2*. The ice at locations *6* and *7* near the snout and at location *1*, high on the northwest side of the glacier, appears to be essentially stagnant and consists of grains of exceptionally large size. Some of these grains are 6 inches or larger in diameter, and their boundaries are more regular than in the active ice. Grains in active ice averaged 1–2 inches in diameter and had apophyses extending irregularly between other grains. Most of the seemingly smaller grains observed in thin-sections proved to be the corners or tapering ends of larger crystals when examined in successive parallel thin-sections about 1 inch apart. Many of the smaller grains in one section were thus shown to be parts of larger grains in the next section. No zones of small grains, possibly indicative of clastic deformation or mechanical grinding, were noted, except for one small area, less than 1 inch square, which did have numerous small grains.

EQUIPMENT AND PROCEDURES

Universal stage.—A universal stage (pl. 2, *B*) with four axes of rotation (comparable to a three-axis stage plus rotation of the microscope stage) was designed and built by Rudolph von Huene, of the California Institute of Technology, according to the author's specifications. This stage, mounted between two 6-inch squares of polaroid, accommodates a $4\frac{1}{2}\times6$-inch thin-section of ice. It was

PLATE 2

A, Method of cutting out ice blocks previous to making thin-sections. Photograph by B. H. Bieler.
B, Top view of universal stage in use on glacier. Photograph by B. H. Bieler.

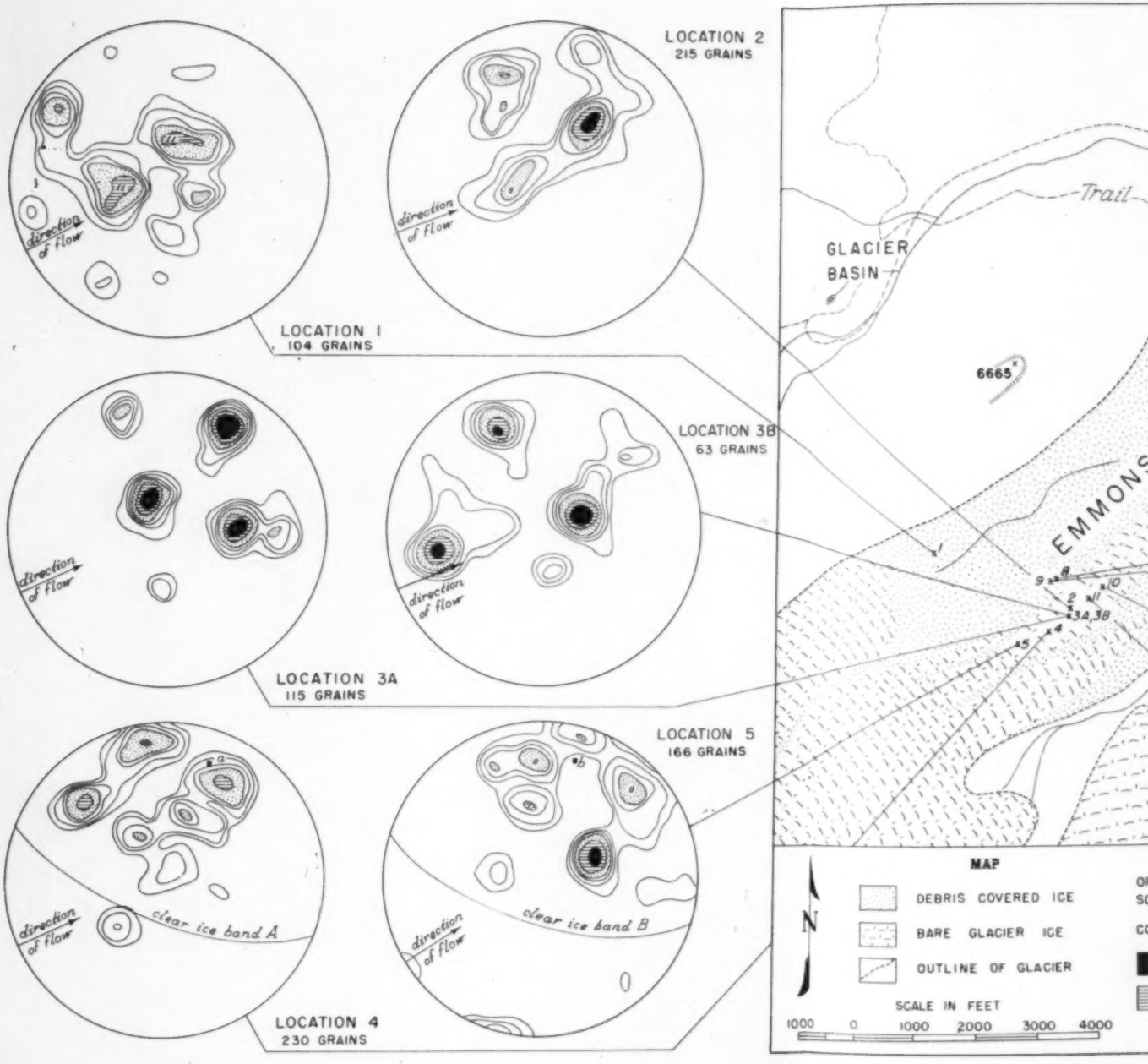
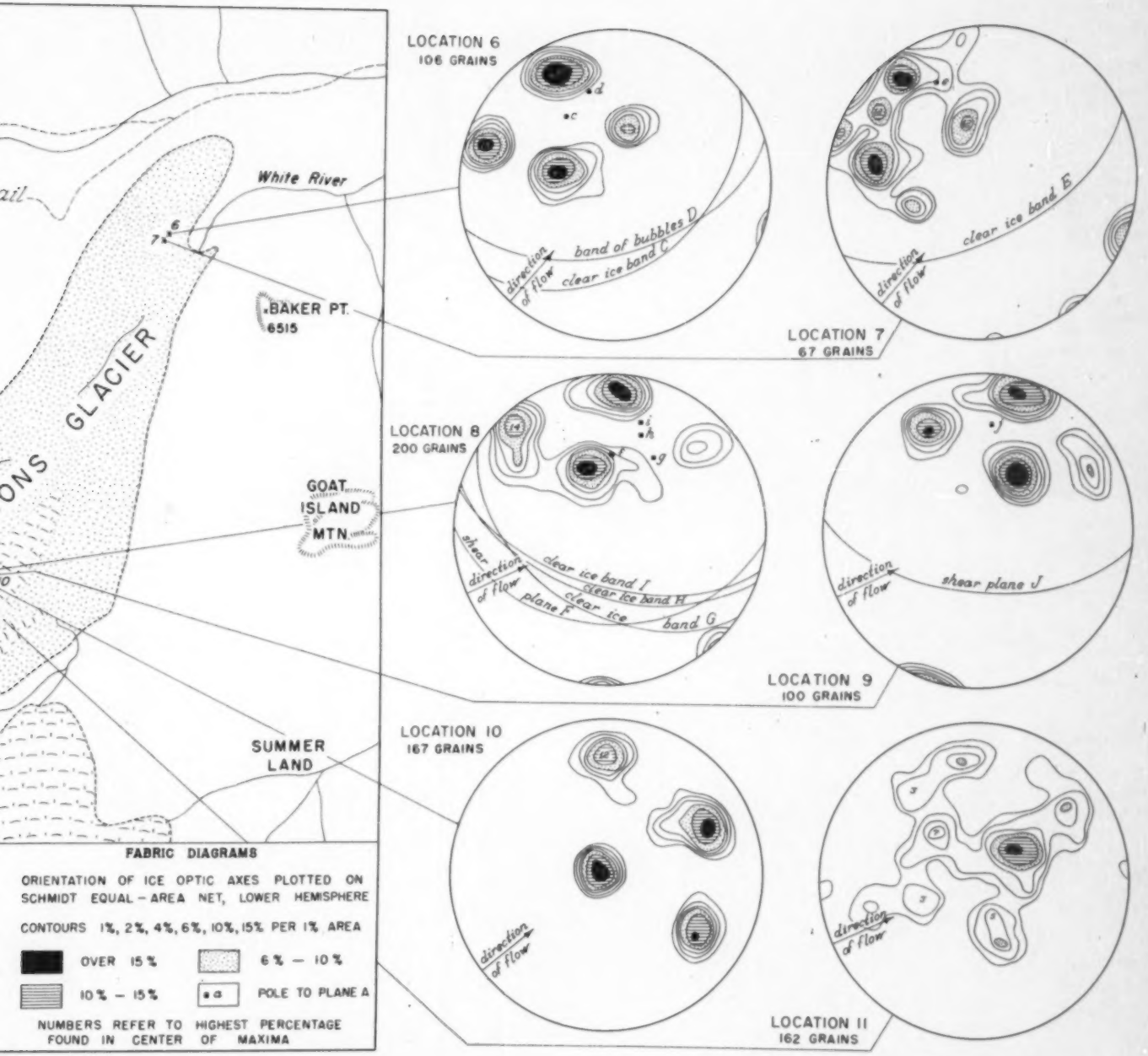


FIG. 2.—Fabric diagrams and outline map of Emmons Gla



Glacier, showing locations where thin-sections were taken

machined out of brass and aluminum to prevent rusting by meltwater. The support for the stage also serves as a carrying case, making a compact and portable field unit.

Thin-sections.—Thin-sections were prepared from slabs of firm, clear ice of known orientation cut from the glacier. After debris and loose, partially melted surface grains had been removed, the surrounding ice was chopped away, leaving a projecting block, from which were cut about four horizontal sections (pl. 2, *A*). These were separated laterally by several inches, so that the same ice crystals were not included in different thin-sections. Each section was cut about $\frac{1}{2}$ inch thick and "ground" to proper thinness by melting with the bottom of a cooking pan filled with warm water. The bottom of this pan had been turned on a lathe to complete flatness and was applied to the ice slab with rotary motion. When the section was about $\frac{1}{8}$ inch thick, it was placed on the stage, as thinner sections broke too easily. The final melting took place in 1 or 2 minutes by contact with the air and glass of the stage. When the ice had melted to a thickness of about $\frac{1}{16}$ inch, extinctions could be measured without appreciable error, although the interference colors were still high. By working rapidly and with the aid of a recording assistant, almost every grain could be measured before the section started to break up and float about on the stage.

The average number of grains per section was 18, but as many as 40 were found in a few sections. Near the glacier's snout, some sections had as few as 3 grains. Fortunately, these were rare, and the average number in areas of coarse-grained ice was 6 per section.

FABRIC DIAGRAMS

Pattern.—The fabric diagrams obtained at eleven locations on Emmons Glacier by plotting the optic axes on the lower hemisphere of a Schmidt equal-area projection are shown in figure 2. Each diagram is in the horizontal plane, and all are oriented in accordance with the map, north at the top. Many of the diagrams have four strong maxima arranged at the corners of a diamond-shaped quadrangle. Some concentrations are as high as 26 per cent in 1 per cent of the area. A few diagrams have other weaker maxima superimposed on the diamond pattern. The data for diagram *3A* (fig. 2) were obtained from sections taken within an area about 14×18 inches on the glacier. The axes plotted in *3B* (fig. 2) were obtained from an area 9×6 inches square, 6–8 inches down the glacier from the *3A* area. The initial expectation was that the grains in the sections from both areas would all have about the same general orientation, but, upon plotting, they proved to be different and have, therefore, been separated. It is surprising that orientations of such distinct differences occur in areas only a few inches apart. Following this discovery, other diagrams, showing optic axes scattered away from the four principal maxima, were replotted by grouping the data from sections in close proximity to one another. Stronger diamond-shaped patterns resulted, even though the number of axes per diagram dropped below the desired minimum. A re-examination of field notes showed that the thin-sections for location *11* had been taken from two areas about 8 inches apart. Figure 3 shows the stronger patterns obtained by plotting the data from these two areas independently.

The thin-sections for location *1* were

taken on successive days from three locations a few feet apart. When these were plotted separately, the four-point pattern appeared more strongly at two of the three spots, but the number of points in each was too small to make a reliable diagram.

Possible origins of fabric pattern.—Ice is known to have a good basal glide plane, and, since Perutz (1940, p. 134) has shown that a preferred orientation of ice crystals is formed along shear planes in glaciers, a possible explanation for the four-maxima pattern might be sought in visible planar structures in the glacier. Shear planes containing rock debris were visible at locations *8* and *9*, and clear-ice bands, usually c.5–1 inch wide and alternating with bands of bubbly ice, were also observed at many locations. The orientation of these planes was measured by Brunton compass. Banding in glaciers is believed to be related to differential movement (Perutz and Seligman, 1939, pp. 347, 356–357). Thin-sections showed that crystal boundaries are independent of the banded structure in Emmons Glacier, for a single crystal commonly crossed two or more ice bands. The debris bands also passed through individual grains without regard to crystal boundaries. This suggests the possibility of recrystallization after shearing, which might destroy any relation between the present pattern and the *s*-surfaces. However, when the poles to these planes were plotted on the fabric diagrams (fig. 2), they fell, for the most part, near the centers of the diamonds formed by the four maxima. The possible significance of this relation is speculated upon as follows:

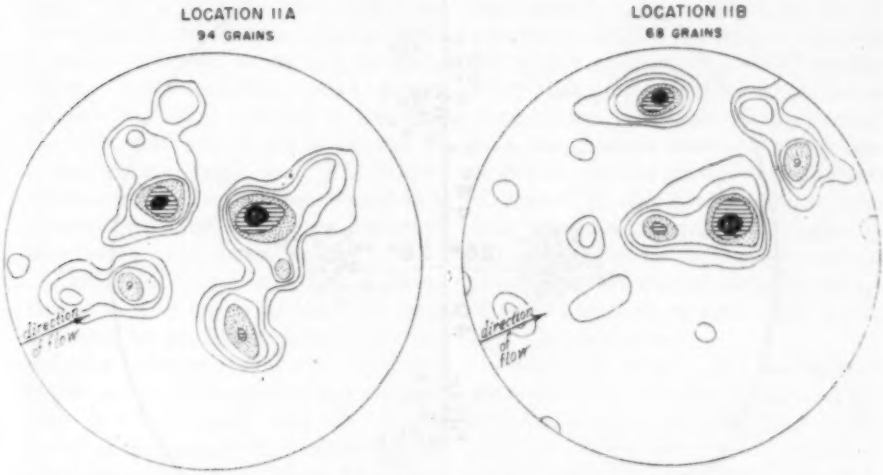


FIG. 3.—The thin-sections at location *11* came from two areas about 8 inches apart. When the data from these two areas are plotted independently, as above, stronger patterns result; compare diagram for location *11* on figure 2.

Let it be postulated that the center of each diamond-shaped pattern is the pole to a shear plane in the ice. On this basis, all the diagrams have been rotated to bring the shear plane into a horizontal position, making the pole to the plane the center of the diagram. The individual

diamond patterns have then been rotated about this pole until their long and short diagonals coincide. The results of these manipulations are plotted in figure 4, where the maxima of the six strongest patterns are represented by dots. The weaker patterns, requiring a more subjective analysis, are represented by crosses. All maxima plotted are numbered according to their locations. A striking uniformity in the angular distance between maxima is immediately apparent.

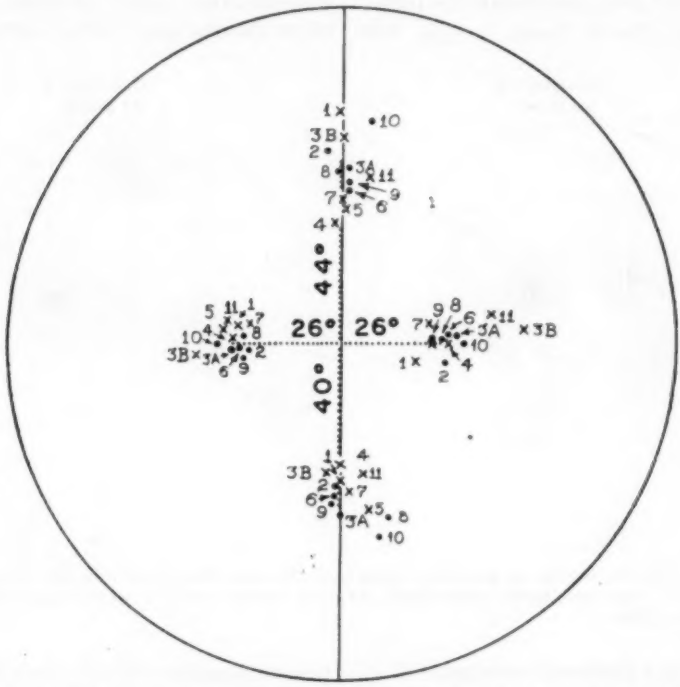


FIG. 4.—Plot of centers of maxima after postulated shear planes have been rotated parallel to horizontal plane, and then all diamond-shaped patterns have been rotated about center until long and short diagonals are parallel. Angular distances between pole to shear plane and maxima shown in degrees.

The angular separation of opposite corners of the six strongest patterns average 84° and 52°, respectively. If the weaker patterns are included, the angles are 80° and 53°, respectively. If these patterns are caused by gliding on planes within the crystal and if only one shear plane in the glacier at each location is responsible for these maxima, then the patterns cannot be attributed to movement on the basal glide plane. Under these assumptions the diamond patterns require two planes with a glide line or direction in each.

The angles between the *c*-axis and the poles of three of the crystal forms commonly found in the hexagonal system are

calculated to be approximately 43° for a $\{11\bar{2}2\}$ face; 28°.5 for a $\{10\bar{1}2\}$ face; and 25° for a $\{11\bar{2}4\}$ face, using the unit cell given by Bernal and Fowler (1933, pp. 523–524) with 12 water molecules. These appear to be possible faces in the structure of ice but not necessarily the most favorable faces for gliding. Usually glide planes are parallel to actual crystal faces, but no good data on actual pyramidal or other faces inclined to the *c*-axis in ice have been found. Barnes (1929, p. 682), however, obtained strong and medium-strong X-ray reflections from planes parallel to $\{11\bar{2}2\}$ and $\{10\bar{1}2\}$ but lists the reflection from a plane parallel to $\{11\bar{2}4\}$ as exceedingly weak. The Miller-Bravais indices for his planes, based on a 4-molecule unit cell, have been converted into the 12-molecule cell.

Using the $(11\bar{2}2)$ and $(10\bar{1}2)$ faces, two glide lines satisfying the conditions indicated by the fabric patterns are (1) the edge between $(11\bar{2}2)$ and the prism $(11\bar{2}0)$ and (2) a line perpendicular to the edge between $(10\bar{1}2)$ and the prism $(10\bar{1}0)$ or the trace of the *c*-axis projected onto the $(10\bar{1}2)$ face. These two lines coincide with the one drawn in figure 4. By interchanging these lines, one could obtain the same pattern, but the glide line on figure 4 would be rotated through 90° in the glide plane to an east-west position. Perhaps an edge between two faces making the proper angles with the *c*-axis can be found to fit the data by rotating the possible glide line 40°–50° on the same figure.

As an alternative hypothesis, the possibility that the fabric patterns were formed by gliding on the basal plane along four shear directions within the glacier is considered. This concept would be supported by a diagram such as the one at location *8*, if the poles of each of the four planes plotted lay in the center of each of the maxima. The fact that they do not may possibly be due to errors in measurement of shear-plane orientation, a difficult procedure in the field. Shearing on these planes would presumably have to occur at different times, and the maxima would be of different ages. Support for this might be drawn from the fact that some of the maxima are weaker than others, possibly being partly destroyed by the later shearing. If this were the proper explanation, it does not seem likely that such consistent angular relations between the shear planes or *s*-surfaces would persist over such widely separated locations as are represented by the diagrams. It also seems unlikely that four planes with consistent angles to one another would form synchronously throughout the glacier. With the exception of location *8*, no indication of four shear planes was seen.

Sources of error.—No hemispheres were used with the thin-sections on the universal stage, and inclinations of the stage of nearly 68° were required to measure axes inclined 45° to the thin-section. At these higher angles, errors seem to have crept in, possibly because there is no means of keeping the line of sight exactly perpendicular to the plane of the stage. Inclinations of the axes of several favorably oriented grains were checked, first, by orienting the optic axis parallel to the line of sight and, second, by bringing it parallel to one of the horizontal axes of the universal stage, which should give the complement of the first angle measured. After being corrected for index difference between ice and air, the sum of these angles totaled about 80° instead of 90°. These errors may have been caused by other factors or combination of factors, such as the difficulty in

locating the exact extinction at high angles. Unusually thick sections appear dark over angles of 20°–30° near true extinction, but, because the sections were allowed to melt until they showed sharp extinction positions, this factor should not have been important. It has been suggested that these errors might arise from ice that had been strained enough to be biaxial with an optic angle of several degrees. It is not known whether this happens with ice as with other minerals, and the ice did not appear to be biaxial on the universal stage. Because the exact error is not known for each measurement and as it probably decreases with decreasing angles, no correction has been made in plotting the data. The effect of the error in axes of high inclination is seen in the shape of contours for certain maxima located about 45° from the center of the diagrams. The splitting of the right-hand maxima at locations *3A* and *4* (fig. 2) are examples.

CONCLUSIONS

Perhaps better explanations can be devised for the fabric patterns found in Emmons Glacier than the ones offered. Further research on deformation of ice is badly needed, but it is hoped that these field data will aid in formulating a clearer conception of the mechanics of glacier flow. The possibility that there may be glide planes in ice other than the well-known basal plane appears worthy of further investigation.

ACKNOWLEDGMENTS.—The work was done under the supervision of R. P. Sharp, and appreciation is expressed for his many suggestions as well as for critical reading of the manuscript. The writer is indebted to Barrie Bieler for assistance in the field and to Rudolph von Huene, who built the universal stage. Thanks are also expressed to A. E. J. Engel and F. J. Turner for suggestions concerning the interpretations of the diagrams. The co-operation and assistance of personnel connected with Mount Rainier National Park is gratefully acknowledged.

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FEATURES OF THE FIRN ON UPPER SEWARD GLACIER ST. ELIAS MOUNTAINS, CANADA[1]

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ABSTRACT

Particular attention is devoted to diagenetic changes within firn, leading to an increase in density and producing structures such as ice layers, lenses, and glands.

Density increases irregularly downward in the Seward firn, with layers of exceptionally low density underlying impermeable horizontal ice masses. The mean rate of density increase with depth in the first 50 feet is 0.0054 per foot, being more rapid—0.01 per foot—in the first 20 than in the last 30 feet, where it is 0.0023 per foot. The rate of density increase in time varies with depth and with the season. Mean density of the 1948–1949 firn layer increased by 0.10 in ten weeks during the summer of 1949, but mean density of the 1946–1947 layer at a depth of 15–18 feet increased only 0.021 in a year. On the upper Seward, a geophysically temperate glacier, the increase of firn density is caused about 25 per cent by meltwater freezing and about 75 per cent by compaction-settling within the zone of annual chilling, the upper 40–50 feet.

Horizontal layers and lenses of ice in firn are formed by freezing of meltwater that has percolated downward and collected above or moved laterally along impermeable horizons. An estimated 20–25 per cent of the Seward firn is converted to solid ice by this process before passing beyond the depth of annual chilling.

Ice glands—crude vertical columns of ice within the firn—are likewise attributed to the freezing of meltwater that has been concentrated along channels extending downward into the firn. The process is pictured as occurring in two stages: first the development of a water-rich condition without much freezing, followed by freezing of the entire mass as a unit, rather than as a process of incremental growth produced by alternate meltwater invasions and repeated freezings along a single channel. As ice layers and lenses are thickest and most numerous near ice glands and extend horizontally from them, a common origin is indicated. These bodies of ice form within the firn during spring when the winter's chilled layer is being ameliorated by the heat given up through freezing of the meltwater that has percolated downward from the surface.

INTRODUCTION

Investigations of the environment and constitution of firn and of its structures and alterations are a logical first step in gaining an understanding of the nature and origin of glacier ice. In addition, firn, like any sedimentary accumulation, records past conditions and events. Firn mantles on glaciers have been extensively investigated in many parts of the world, but they have been singularly neglected in North America. Therefore, the opportunity afforded by "Project Snow Cornice," an endeavor directed by Walter A. Wood, under sponsorship of the Arctic Institute of North America, to study firn in a remote area along the Alaska-Yukon border, was most timely.

PHYSICAL SETTING

Seward Glacier consists of two parts, a large intermontane accumulation basin of approximately 515 square miles, herein designated "upper Seward Glacier," and a narrow outlet stream, 20 miles long. This outlet glacier, the lower Seward, discharges with high velocity into Malaspina Glacier, a piedmont ice sheet on the southern Alaska coastal plain (fig. 1). Attention is confined to the upper Seward, lying wholly within the Yukon Territory of Canada in the heart of the St. Elias Mountains. It is almost completely encircled by high peaks, culminating to the southwest in Mount St. Elias, 18,008 feet, and to the northwest in Mount Logan, 19,850 feet. The gently undulating, firn-mantled surface of upper Seward Glacier lies at elevations of 5,000–7,000 feet, 2,000–4,000 feet above the developed firn limit.

Although the upper Seward is a scant 50 miles north of the Gulf of Alaska, the high mountain barrier to the south gives a meteorological environment more con-

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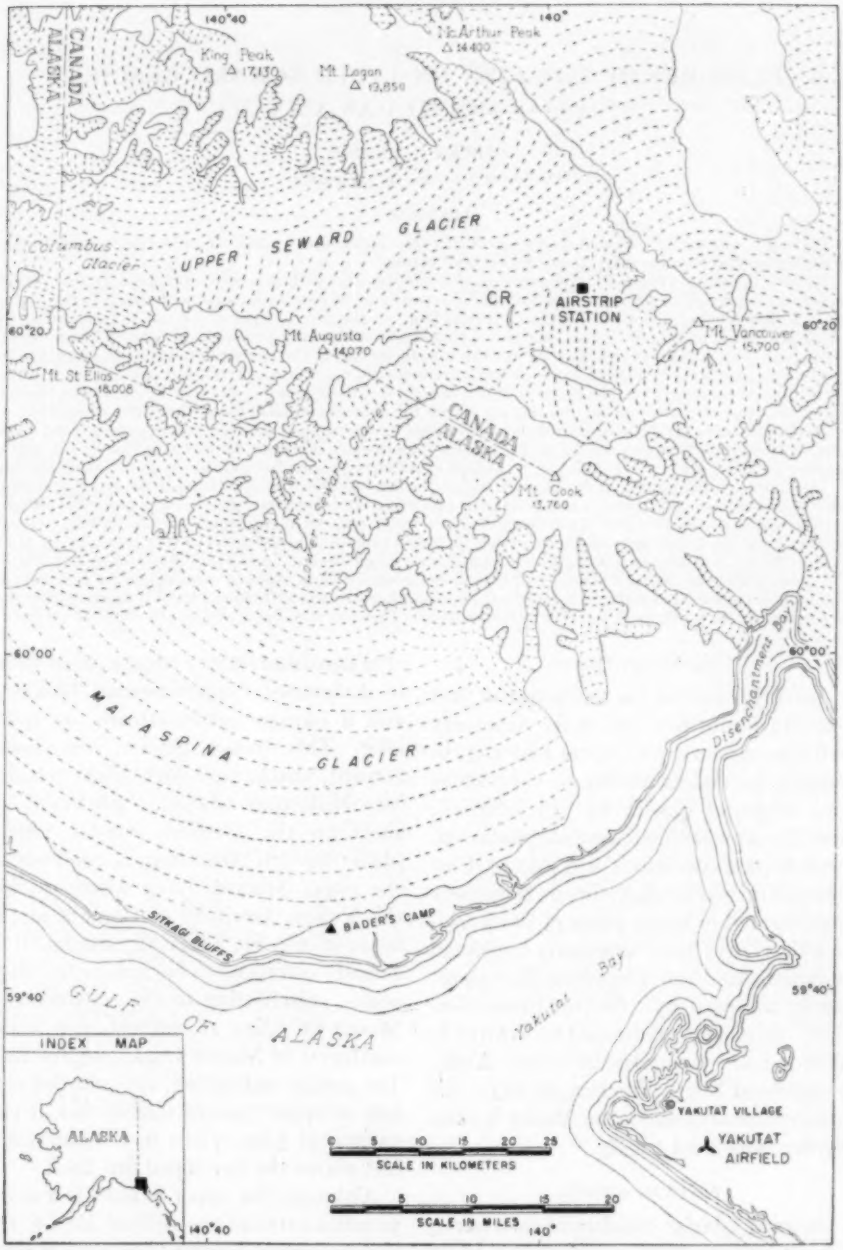


FIG. 1.—Map of Seward-Malaspina Glacier system. *CR*, Crud Ridge

tinental than coastal. The summers have many nearly cloudless days, with brilliant sunshine and maximum air temperatures in the middle or high fifties Fahrenheit. Temperatures fall to the middle and low twenties Fahrenheit on nearly every clear night, but freezing may not occur at all on clouded nights. Some wet snow and rain fall during summer, but monthly precipitation during July and August does not usually exceed 0.5–1.5 inches.

Studies of firn centered at the Airstrip Station (60°23′ N. lat., 139°15′ W. long., elev. 5,875 feet) and its environs within a radius of 6 miles on the eastern part of upper Seward Glacier. At the Airstrip a firn mantle, of 50–60 feet in estimated thickness, overlies approximately 2,150 feet of ice as determined by seismic reflections. The upper Seward is geophysically temperate (Ahlmann, 1948, p. 66), with the firn and ice at the pressure-melting temperature except for a surface layer 40–50 feet thick which is chilled to subfreezing temperatures in winter (Sharp, 1951*a*). Field work covered the periods of July 10 to August 29, 1948, and June 15 to August 29, 1949.

TERMINOLOGY

For the sake of understanding and consistency, a few frequently used terms are defined.

Old snow is granular, settled snow which has not yet attained a density of 0.4.

Firn is compacted granular snow with a density of 0.4 or greater. It is distinguished from solid ice by being pervious and having a density no greater than 0.82–0.84.

Ice is a solid, relatively impervious mass of ice crystals with density of 0.84 or greater.

Iced firn is a mixture of ice and firn.

Ice gland is a crudely shaped vertical column of ice or iced firn.

Firnification embodies the processes converting snow into solid ice.

FIRN DENSITY

EQUIPMENT AND PROCEDURE

Determinations were made by weighing a sample of known volume and calculating the bulk density. Several methods of sampling and different types of equipment were tried. Best results were obtained by using the volume of the hole from which a specimen was taken, in the calculation, rather than the volume of the specimen itself. This proved particularly true in hard, compact firn that was broken by sampling. A cylindrical steel tube about 70 cm. long, with an outer diameter of 3.8 cm., proved most satisfactory as a sampler, and details of its design and use are given in the caption of figure 2. Core barrels of larger diameter gave no greater accuracy and had the disadvantage of sampling parts of adjacent firn layers of different densities. In 1948 volumes as small as 306 cm. were used because of a short core barrel, but in 1949 all holes from which samples were taken had a volume of 500 cc.

The sampler used was not sturdy enough for solid ice, so holes of known diameter and length were bored into ice with a carpenter's brace and a large-diameter bit. Cuttings were collected and weighed, and the bulk density calculated from the known volume of the hole. Although this is simpler than the immersion technique used by Moss (1938, p. 221), Seligman (1941, p. 302), and Wade (1945, pp. 162–163), the results obtained were not wholly consistent, and the procedure needs further study and refinement for use in solid ice.

Because the firn is horizontally stratified (pl. 1, *B*), all samples were taken horizontally from the walls of pits in order to obtain homogeneous specimens. Vertical samples suitable for calculation of total water equivalents show nothing of the density differences between firn layers, a matter of major interest here. Newly dug pits are best for density determinations, and the walls of pits only a few days old must be cut back a foot or more to unaltered firn. For this reason, comparative density determinations cannot be made in exactly the same places on different dates, and variations are unavoidably introduced by vertical and horizontal inhomogeneities within the firn. The conformal symmetry of density curves made on different dates (fig. 3) shows that these differences need not obscure the major trends, although they may have significance in comparisons of individual values.

The amount of liquid water within thawed firn at the time of sampling is a matter of concern in density determinations. This water is of two types, fixed and transitory. The fixed liquid water can properly be considered a part of the firn, as it is relatively constant and, when within 50 feet of the surface, will be frozen during the winter. However, transitory liquid water varies from hour to hour, depending on the position and volume of the daily meltwater wave moving downward through the firn. Liquid-water determinations by Bader's (1948, pp. 355–357) temperature-depression method suggest a variation in transitory liquid water not exceeding 7–8 per cent of the total firn mass. Such variations could introduce maximum errors of 0.04–0.05 in the average density value of firn, although most errors from this source probably do not exceed 0.02. In comparative density studies the liquid-water in-

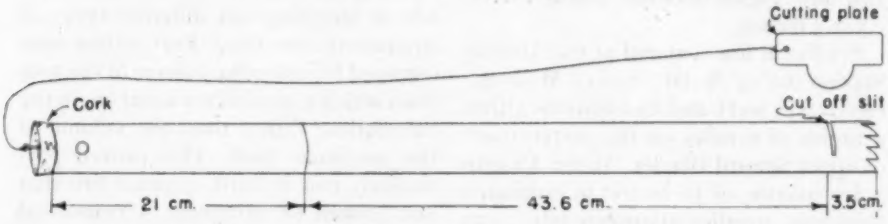


FIG. 2.—Coring device for taking density samples of firn. The teeth are shaped to force the cuttings into the barrel. The sharpened cutting plate is inserted in the slot at 3.5 cm., and the irregular broken end of the core is discarded. The error introduced by this procedure involves only the volume increase within the discarded material and is insignificant. The mark encircling the barrel at 43.6 cm. indicates a hole volume of 500 cc. between the slot and the mark, as calculated from the outer diameter of the tube. The cork prevents loss of material from the outer end of the barrel.

PLATE 1

A, Ice layers and lenses in crevasse at depth of 40 feet near outlet of upper Seward Glacier. Crevasse is 2–3 feet wide.

B, Horizontal stratification in firn near northeast corner of upper Seward Glacier.

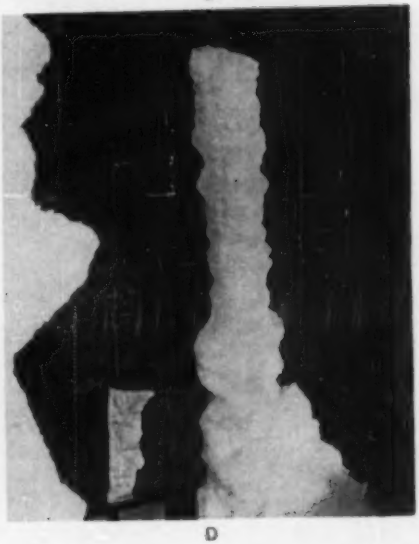
C, Two ice glands exposed by excavation. Slight doming of firn by differential compaction shown above larger gland with horizontal projections.

D, Close view of smaller gland shown in *C*. Horizontal layering is typical. Height about 3 feet.

A

B

C



D

Layering and ice glands of upper Seward Glacier

A

B

Ice glands in firn of upper Seward Glacier

fluence can be largely eliminated by taking corresponding samples at the same time of day and under reasonably similar ablation conditions. For more accurate results the amount of transitory liquid water would have to be measured and appropriate allowances made. As far as is known, this has never been done in any study of firn density.

DENSITY VALUES IN THE SEWARD FIRN

Samples were taken from the walls of pits, either freshly dug or cut back to unaltered material. Samples were spaced from a few inches to 1 foot apart, depending on visible inhomogeneities in the firn. In 1948 a pit 50 feet deep was dug, and densities were measured to that depth. In 1949 attention was devoted chiefly to recording density changes occurring within the 1948–1949 firn layer during the summer, but measurements were also made in the 1946–1947 and 1947–1948 firn layers for comparison with values obtained in 1948. Densities are calculated to the nearest hundredth, as that is the least significant figure permissible with the method and equipment used and the variable influence of liquid water may introduce errors exceeding this degree of accuracy. All figures are for bulk density.

The surface material at the Airstrip Station had already attained the state of firn at the earliest dates of observation, specifically July 10, 1948, and June 15, 1949. In 1948 all material sampled to a depth of 50 feet, aside from ice layers, qualified as firn; but in 1949 old snow of density 0.37–0.39 (fig. 3) was found beneath the denser surface material. This snow was converted to firn by July 6, 1949. Densities to a depth of 50 feet as determined in 1948 ranged from 0.47 to 0.85. Location and spacing of samples control in some degree the details of the density curve (fig. 4), but the irregularities recorded cannot exceed, and may be somewhat less than, the actual differences, because absolute maximum and minimum densities may be missed in sampling. Despite the irregularities, this curve shows the gradual density increase normally expected with depth. The mean rate of increase is 0.0054 per foot, but density increases more rapidly in the first 20 feet, 0.01 per foot, than in the last 30 feet, 0.0023 per foot Also noteworthy are the layers of exceptionally low density beneath extensive ice layers Although not universal, this arrangement is consistent enough to be significant in respect to processes which increase firn density.

Densities within the 1948–1949 firn layer (fig. 3) are likewise progressively greater with depth and have relatively lower values beneath impervious ice layers. A progressive density increase is also recorded as the season advances (fig. 3). Mean density for the 1948–1949 firn layer increased from 0.44 on June 20 to 0.54 on August 29. Extreme values were 0.37–0.50 on June 20 and 0.45–0.64 on August 29. The rate of increase within the 1948–1949 firn layer was 0.10 in 10 weeks. No increase of mean density was detected within the 1947–1948 firn layer between 1948 and 1949, since it was 0.51 in both years; but the 1946–1947 firn layer showed an increase from 0.514 to 0.535 during the same period. This rate

PLATE 2

A, Small knobs are tops of ice glands exposed on the firn surface by ablation, July 16, 1948.
B, Lines of ice gland knobs extending downslope, August 19, 1949.

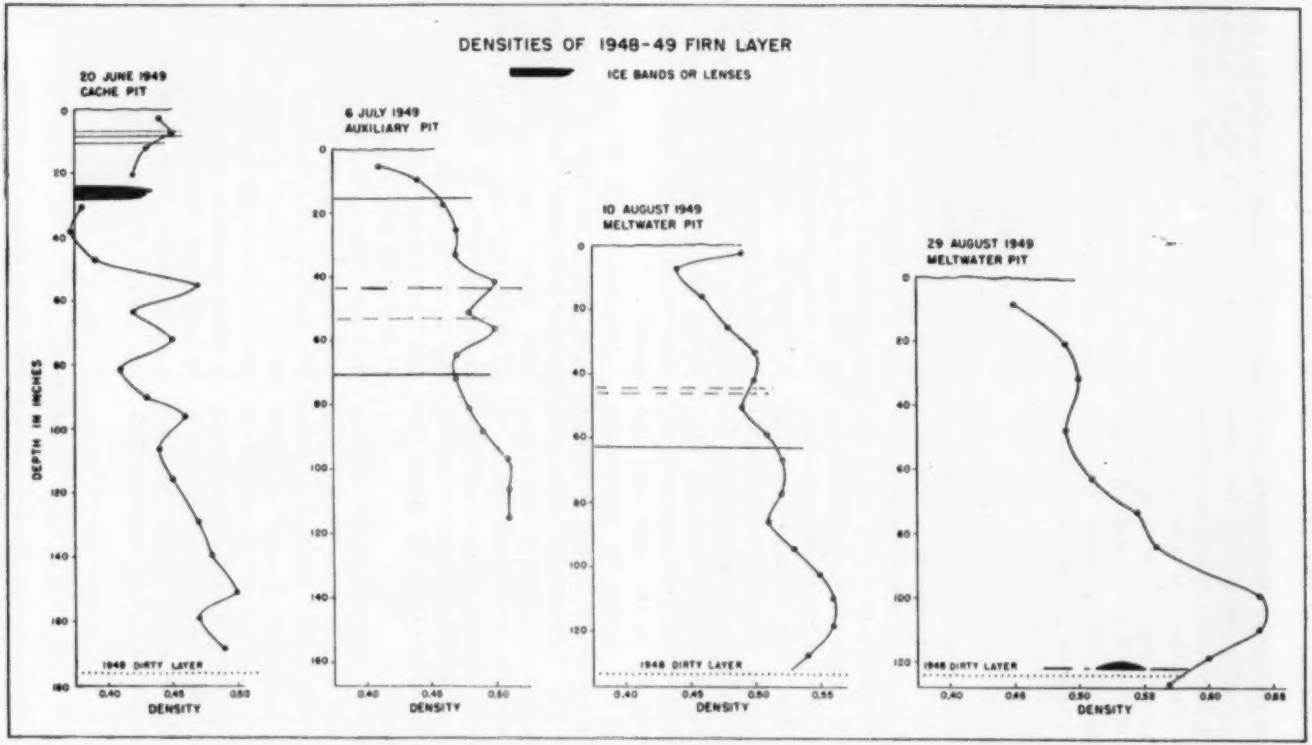


FIG. 3.—Bulk density of 1948–1949 firn layer at different times during the summer of 1949 in various pits at the Airstrip Station

of increase, roughly 0.02 per year, is about one-fifth that recorded in the 1948–1949 layer during the summer of 1949. The rate obviously decreases rapidly with depth and is also probably much slower during winter. Failure to record an increase within the 1947–1948 layer may have been due to sampling errors, which missed some of the denser layers in 1949, or to the liquid-water influence. All the above figures apply to firn and not to the bodies of ice developed within it.

In early August, 1949, mean density in the 1948–1949 firn layer was 0.510; in the 1947–1948 layer, 0.511; and in the 1946–1947 layer, 0.535. This rate of increase is somewhat more rapid in time than that reported from West Spitsbergen (Ahlmann, 1935*b*, p. 102) and Greenland (Sorge, 1933, p. 339) but less rapid than in the Alps (Seligman, 1941, p. 308). With respect to depth, it is less than in North East Land (Glen, 1941, p. 143) and Antarctica (Wade, 1945, pp. 168, 170–172). These contrasts probably reflect differences in thermal and meteorological environments and rates of accumulation.

The density of ice masses within the firn, as determined in 1948 by the procedure previously described, ranged from 0.86 to 0.90, which is comparable with measurements in other areas (Moss, 1938, p. 227; Etienne, 1940; Wade, 1945, p. 165).

CAUSES FOR INCREASE OF FIRN DENSITY

General statement.—Firnification has been a subject of study for over a century (Rendu, 1874, p. 62), and notable contributions to the problem have been made in the last fifteen years by Scandinavian (Sverdrup, 1935, pp. 72–80; Ahlmann, 1936, pp. 63–73; Ahlmann and Thorarinsson, 1938, p. 415; 1939, pp. 42–55), British (Hughes and Seligman, 1939, pp. 640–645; Glen, 1941, p. 145), and Swiss scientists (Bader, 1939). Their work shows that the principal causes of

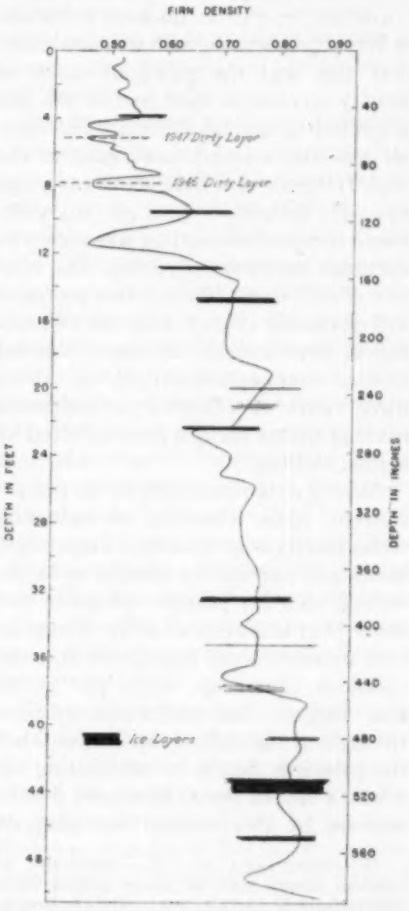


FIG. 4.—Bulk density of firn in deep pit at Airstrip Station, 1948.

firn density increase are: (1) freezing of meltwater percolating down from the surface and (2) compaction-settling. Sublimation, liquid transfer, regelation, and crystal growth are considered secondary

activities contributing primarily to compaction-settling.

Freezing of surface thaw water percolating into firn was investigated by Sverdrup (1935, pp. 71–80) and Ahlmann (1936, pp. 63–73) on Isachsen's Plateau in West Spitsbergen, with the conclusion that this was the principal cause of density increase in that part of the firn subjected to annual chilling.[2] The matter was also studied thoroughly in the Alps (Hughes and Seligman, 1939, pp. 640–645; Seligman, 1941, pp. 303–308), where compaction-settling was shown to outweigh meltwater freezing. The relative effectiveness of these two processes will obviously change with the climatological environment. On upper Seward Glacier compaction-settling is about three times as effective as meltwater freezing within the firn layer affected by annual chilling.

As long as temperatures in the firn are subzero, some refreezing of meltwater seems inescapable. Indeed, ice layers and lenses and vertical ice glands, to be described shortly, provide adequate evidence that this occurs locally. Temperature measurements and theoretical calculations (Sverdrup, 1935, pp. 71–76) also suggest that meltwater refreezes throughout the bulk of the firn, as this is the principal means of eradicating the winter's chilled layer. However, density increase by this method has temporal limitations, for it cannot occur without surface thawing and ceases when the winter's chilled layer is ameliorated. On most temperate glaciers, meltwater freezing takes place chiefly in late spring and early summer. It also occurs during fall and winter when a new chilled layer develops, but the amount of water held in the firn at this time is small (Sverdrup, 1935, p. 74). Meltwater freezing also has vertical limitations determined by the thickness of the annual chilled layer. On upper Seward Glacier the chilled layer is 40–50 feet thick (Sharp, 1951*a*). Because the annual accumulation surplus averages about 60 inches of firn (Sharp, 1951*b*), each annual firn layer is subjected to freezing of meltwater for about a decade.

By contrast, compaction-settling acts the year around and has no depth limitations. Its effectiveness increases with temperature (Bader, 1939, p. 49) but probably decreases with depth beyond a certain point, owing to an asymptotic approach of the firn density to the limiting density of solid ice. In areas where meltwater is lacking, compaction-settling and its contributing processes may be the sole mechanism of density increase (Sorge, 1933, p. 339; Wade, 1945, pp. 165–166; Howard, 1948, p. 921).

Data from the Seward firn.—Many layers of firn lying just beneath horizontal ice masses (figs. 3 and 4) have exceptionally low density because they have been deprived of their proportionate share of meltwater by the overlying, relatively impervious ice bodies. The fact that not all firn layers just below ice masses have low density can be attributed to leaks or to lateral movement of water from some other source. A crude calculation of the minimum contribution of meltwater to the increase of firn density is possible from the following data.

[2] Sverdrup's (1935, pp. 74–75) instructive calculations assume that the density increase under consideration is due solely to meltwater freezing, although compaction-settling undoubtedly plays a part in Spitsbergen as elsewhere. Even ignoring compaction-settling, his calculated values exceed the observed values by as much as 0.036 and by even greater amounts if it is assumed that some fixed liquid water is held in the firn to be frozen by next winter's chilling. This suggests that the observed density values may be too low, possibly because the firn was disturbed during sampling, as described above. Since the method used to determine densities is not given, this possibility cannot be evaluated.

The density of firn underlying some ice layers is 0.07 below the average for adjacent firn layers. If old snow of density 0.30 is assumed to have been the original surface material and if the average density of the firn adjacent to the low-density layers is 0.53, then freezing of meltwater has supplied not less than 30 per cent of the increase from 0.30 to 0.53. The amount contributed by meltwater may actually be larger, because it cannot be assumed that the low-density material is completely deprived of meltwater.

Changes of density within the 1948–1949 firn layer during the summer of 1949 are instructive. The first curve of figure 3 represents densities on June 20, 1949, when the material below a depth of 44 inches was still "frozen" and some of it qualified as old snow with density 0.37–0.39. A surface layer 10–20 inches thick consisted of grains averaging about 1 mm. in diameter, but the rest of the grains to the bottom of the 1948–1949 layer were 0.25–0.5 mm. in diameter. The average density for the entire layer was 0.44. By July 6, 1949, this annual layer had been warmed to 0° C., its average grain size in the upper 9 feet had increased to 0.5–0.75 mm., and the average density of the upper 9 feet was 0.48. Owing to an oversight, the sampling pit was not extended to the bottom of the 1948–1949 layer.

The density change between June 20 and July 6 was the largest recorded for any equivalent time interval throughout the summer. Initially, it was thought that the freezing of meltwater required to raise the firn to 0° C. was the principal cause of this increase, but calculations based on recorded firn temperatures show that meltwater freezing contributed not more than 10 per cent of the increase. Because this marked density change occurred during the time that the firn attained a thawed condition, it is suggested that thawing permitted the firn grains greater freedom of movement, and the firn experienced an exceptional amount of compaction-settling at this time. As a general principle, it would seem that firn must undergo an exceptional increase in density as it passes from the frozen to the thawed condition.

Density increase continued within the 1948–1949 firn after it reached 0° C. early in July (fig. 3). This must have resulted from compaction-settling, produced in part by a shifting-about of the individual grains. Grain movements may not occur easily where meltwater is lacking (Wade, 1945, p. 166), but studies of firn fabric on a temperate glacier (Perutz and Seligman, 1939, pp. 344–345) show that a preferred crystallographic orientation established near the surface is destroyed as the firn becomes more deeply buried, presumably because the grains have moved. A slow growth of grains by vapor and liquid transfer arising from differences in size and possibly from pressure differences may also have had some effect, for a slight increase in grain size seems to have occurred after the Seward firn attained 0° C.

Attempts to measure compaction in the Seward firn during summer proved fruitless because high air temperatures caused the compression plates to undergo independent movement by melting. However, doming of firn layers over ice glands (pl. 1, *C*), as observed in walls of pits and crevasses, shows that compaction-settling occurs. A crude measure of the compaction-settling that had occurred within the lower half of the 1948–1949 firn layer by June 20, 1949, was obtained accidentally from some toboggan sleds left in an upright position in a cache on the 1948 surface. These sleds were deeply buried within the 1948–1949

firn layer, and one of them, originally 84.5 inches long, was crushed and shortened to a length of 73.25 inches by the overlying 90 inches of snow and firn. Shortening of the sled by 11.25 inches gives a minimum measure of settling within the lower 84.5 inches of the

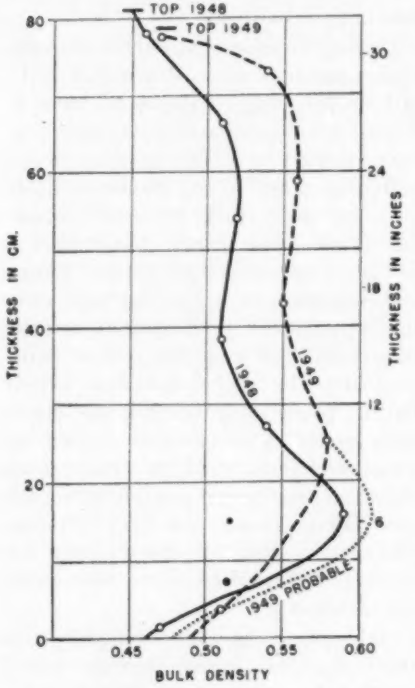


FIG. 5.—Curves of bulk density in 1946–1947 firn layer as determined in the summers of 1948 and 1949.

1948–1949 firn layer between the time the sled was first completely buried—probably in January or February, 1949—and June 20, 1949. This establishes a minimum compaction of 13 per cent in 5 or 6 months.

None of these observations measures the ratio between compaction-settling and meltwater freezing in producing the observed density changes, but a crude approximation is obtained from density changes within the 1946–1947 firn layer between 1948 and 1949. In 1948 average density of the layer was 0.514. According to measurements made in 1949, it was 0.535, if the density curve is extrapolated as shown by the dotted part in figure 5. Fortunately, determinations in both cases were made during the low phase of the daily meltwater wave, so variations in the amount of transitory meltwater probably did not introduce significant errors.

In 1948 the thickness of this layer was 32 inches, and in 1949 it was 31 inches. These measurements are crude and not wholly comparable, as they could not be made in exactly the same places in successive years, but they represent averages of several figures for each year, and the difference of 1 inch is consistent. If the data are accepted, a 3.1 per cent decrease in volume has been accompanied by a 4.1 per cent increase in density. Accordingly, 75 per cent of the density increase would be due to compaction-settling and 25 per cent to freezing of meltwater. This estimate makes no allowance for increase in the amount of solid ice in the form of layers, lenses, and glands, and the 3 to 1 ratio of compaction-settling over meltwater freezing applies only to firn at a depth of 15–18 feet. Here the degree of chilling in winter is about one-third that near the surface, and the amount of meltwater frozen in spring is less, although the higher specific heat of the denser firn is a partly compensating factor.

As an approximation, the density increase occurring within the uppermost 40–50 feet of the firn on upper Seward Glacier is attributed about 25 per cent to meltwater freezing and about 75 per cent to compaction-settling. This is based in part upon data reported above and upon

calculations[3] based on firn temperatures. The "icification" of firn, represented by discrete layers, lenses, and columns of ice formed by freezing of meltwater, is not included in the estimate.

FIRN STRUCTURES

Firn is essentially a sedimentary accumulation, consisting of successive layers of material precipitated directly from the air or deposited by the wind, and, like sediment, it subsequently undergoes diagenetic changes. Firn is gradually subjected to an increasing degree of "load" metamorphism and eventually to "flow" metamorphism, which alter and may even destroy the original primary structures. Because firn and ice are fundamentally monominerallic rocks, they can be studied by the usual procedures of stratigraphy, sedimentation, and petrography. The environment and conditions of deposition of the original material—snow—and the subsequent diagenetic changes can be worked out more easily than for most rocks because they occur rapidly and close to the surface. In this study attention is focused chiefly on features produced by diagenetic changes.

ICE LAYERS AND LENSES

Description.—The walls of crevasses and excavations in upper Seward Glacier expose many horizontal layers and lenses of bluish ice within white granular firn (pl. 1, *A*). Their thickness ranges from a feather edge to 1 foot, but it is seldom constant, for swelling and pinching caused by undulations in the upper surface are nearly universal. The bottoms for the most part are flat, as also noted in other areas (Ahlmann, 1935*a*, pp. 32–34; Glen, 1941, p. 143), and the lower contact of an ice mass is usually sharp, with the upper contact more gradational.

The centers of some lenses underlie the edges of ice layers or lenses higher in the section (fig. 6, *B*), and in places closely spaced ice layers pass laterally into a single thick ice lens (fig. 6, *A*). Lenticularity and discontinuity are characteristic of horizontal ice bodies at shallow depths near the Airstrip Station, but lenses tend to be localized along specific horizons (fig. 6, *G*). Most ice masses, aside from the vertical glands to be described shortly, are essentially horizontal, although some have been gently deformed by differential compaction.

Near the outlet from upper Seward Glacier, east of the south tip of Crud Ridge (fig. 1), horizontal ice bodies exposed in walls of crevasses are thicker and of greater lateral extent than at the Airstrip Station. Little trouble is experienced in correlating these ice layers from one crevasse to another (fig. 7). With increasing depth at the Airstrip Station the ice bodies also become more numerous and thicker and have greater lateral continuity (fig. 4). A high degree of correlation in size and spacing of ice layers and lenses was demonstrated by Miller and Leighton in two crevasses about 75 yards apart at this site (fig. 8). In addition, a prominent icified horizon in the 1947–1948 firn layer has been

[3] If the projected minimum-temperature curve in Seward firn for 1949–1950 (Sharp, 1951*a*, fig. 6), with a minimum of about −13° C. and an estimated depth of 40 feet, is accepted as typical, the density increase within the uppermost 40 feet of the firn is calculated to be 0.138. This assumes an average thickness of 5 feet for each annual firn layer, with the layer at the bottom having experienced chilling for 8 successive years, the layer next to the bottom for 7 years, and so on. Densities used in calculating specific heat of the firn at various depths were taken from fig. 4. An estimated 20 per cent of the 40 feet of firn has been converted to solid ice in the form of lenses, layers, and glands by freezing of meltwater, and an allowance must be made for this, as it reduces the amount of freezing available for increasing the firn density. When so corrected, the density change in firn resulting from meltwater freezing is 0.099, or about 25 per cent of the total change of 0.40.

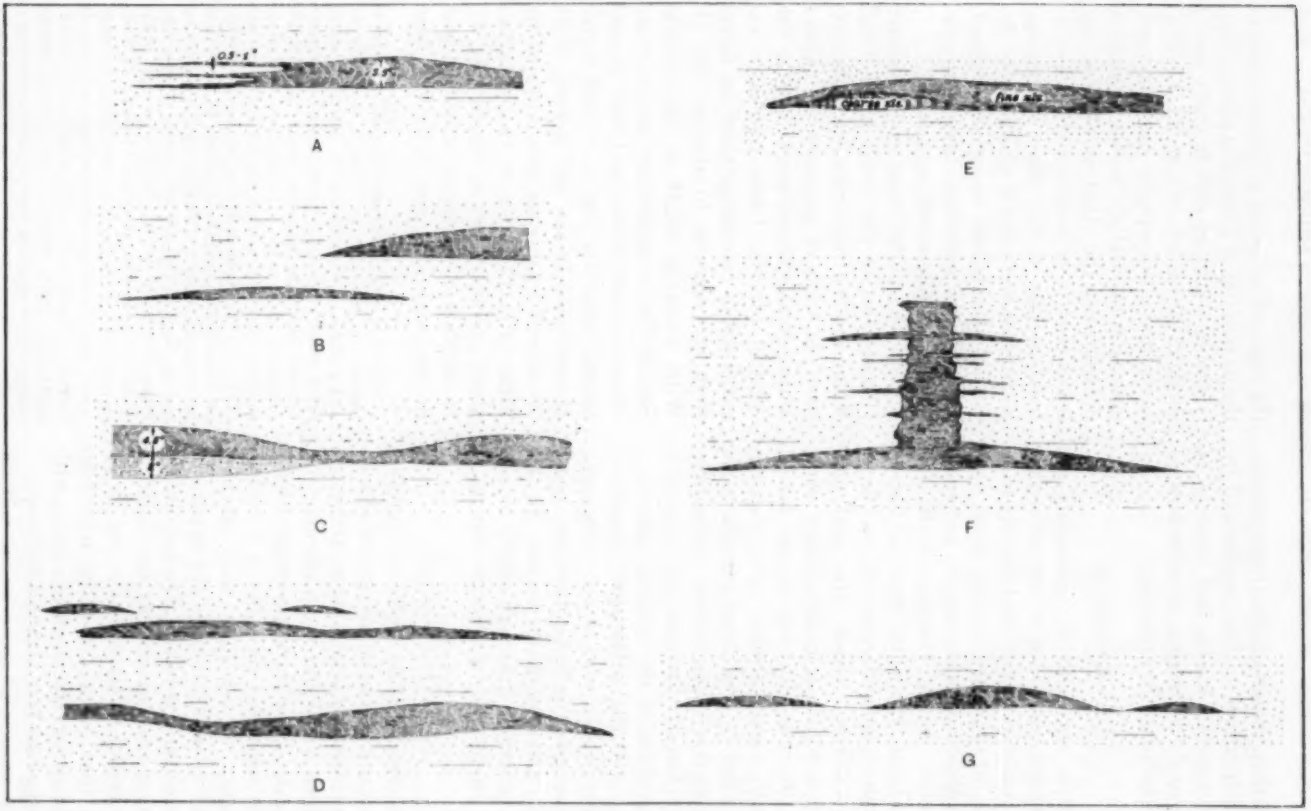


FIG. 6.—Sketches of horizontal ice bodies in firn

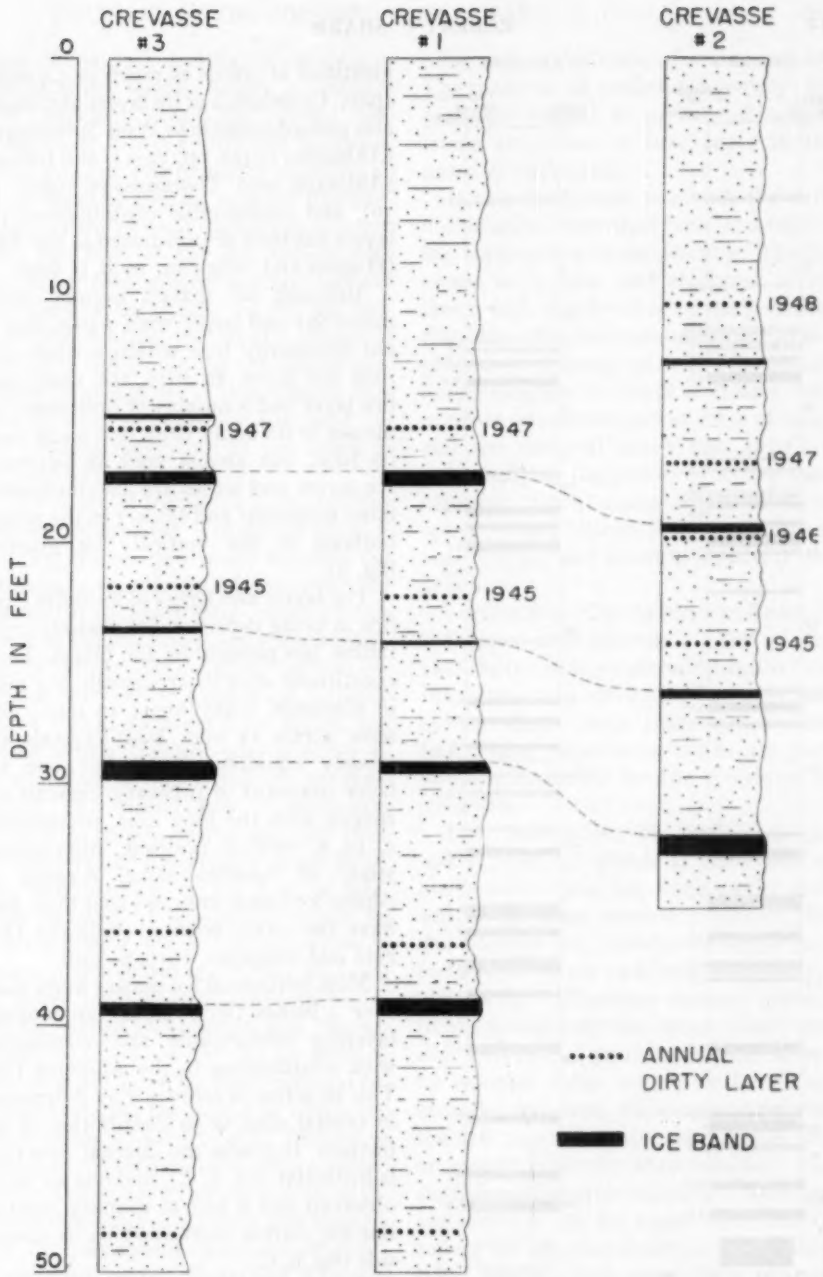


FIG. 7.—Correlation of prominent ice layers and annual dirty layers exposed in walls of crevasses near outlet of upper Seward Glacier.

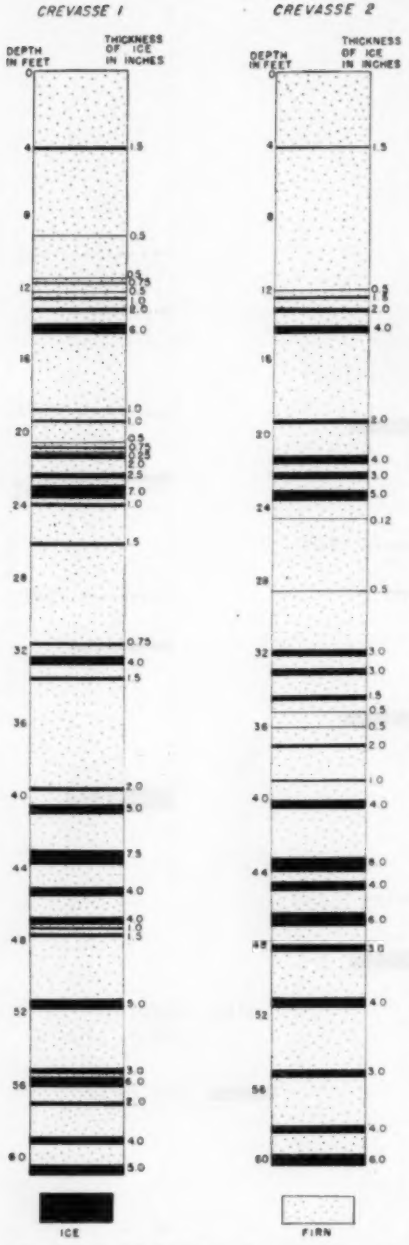


FIG. 8.—Correlation of ice layers and lenses in firn exposed in walls of two crevasses 75 yards apart at Airstrip Station; data from Leighton and Miller.

identified at points as much as 5.5 miles apart. Correlation of ice layers and lenses also proved possible in West Spitsbergen (Ahlmann, 1935*a*, pp. 32–33) and Iceland (Ahlmann and Thorarinsson, 1939, p. 56), and considerable continuity of ice layers has been demonstrated in the Alps (Hughes and Seligman, 1939, p. 640).

Although ice masses become more numerous and larger with depth, this is not necessarily true within a single annual firn layer. In 1949, the 1948–1949 firn layer had a number of horizontal ice masses in its upper third and a few near its base, but almost none in between. Ice layers and lenses are also invariably more numerous and thicker in the neighborhood of the vertical "ice glands" (fig. 9).

The layers and lenses of ice differ from firn in being denser, more coarsely crystalline, less permeable, and bluish. Their constituent crystals are usually 2–5 mm. in diameter, many are 5–10 mm., and some attain 15 mm. Most crystals are crudely equidimensional, but one ice layer consisted of prismatic crystals arranged with the long axes, presumably *c*, in a vertical position. Microscopic study of equidimensional crystals in Alpine ice bands indicates that they, too, have the *c*-axes oriented vertically (Perutz and Seligman, 1939, p. 346).

Most horizontal ice masses more than 1 or 2 inches thick display an internal layering conformable and continuous with stratification in the adjoining firn. This structure is expressed by differences in crystal sizes or in distribution of air bubbles. In places the internal structure is lenticular (fig. 6, *E*), and one ice layer observed had a lens of coarsely crystalline ice, convex downward, on its underside (fig. 6, *C*).

Origin.—Ice layers and lenses have been attributed primarily to freezing of meltwater that has percolated downward

from the surface and has been concentrated along specific layers in subzero firn (Sverdrup, 1935, pp. 78–80; Ahlmann, 1936, p. 65; Perutz and Seligman, 1939, p. 347). The following relations on upper Seward Glacier support this concept: (1) the general increase in number and thickness of the ice masses with depth, (2) their greater size and number near vertical ice glands, (3) the flat bottoms and undulatory upper surfaces of ice layers and lenses, (4) the lateral continuity of structures in these bodies with stratification in adjoining firn, and (5) the offset arrangement of ice lenses in vertical section.

The mode of formation is pictured as follows: Meltwater formed by surface thawing moves downward through the underlying, pervious, subzero firn. For reasons to be given later, much of this water is believed to move most rapidly and in largest volume along vertical columnar channels. It spreads out laterally above the more impervious layers and crusts within the firn and possibly to some degree within especially pervious layers. The lateral flow of meltwater eventually ceases, and conduction from the surrounding subzero firn causes freezing, thus forming an ice layer or lens. This freezing by conduction must be relatively slow, for in Iceland layers of water-saturated firn at 0° C. lay adjacent to layers of firn with temperatures as low as −4°.5 C. (Ahlmann and Thorarinsson, 1939, pp. 46–47). An ice layer or lens need not grow by daily external increments during the period in which the firn is being warmed to 0° C. Instead, a layer may receive meltwater for a period of days before the supply is cut off or diverted into some other path and freezing sets in. Subsequent invasions of meltwater may occur, saturating the firn above and laterally beyond such ice masses and thus adding to their thickness and extent when freezing occurs. Fluctuations of meltwater supply are probably related to periods of surface freeze and thaw or to changes in the paths of circulation.

If this mechanism is sound, the various relations described have the following explanation or significance. The increase in number and thickness of ice layers with depth is due to the fact that the deeper firn has been subjected to the process for a longer period. This has its limitations, for the depth of annual chilling does not exceed 50 feet, here at least, and the material below that depth is perennially at the pressure-melting temperature. Each annual firn layer on upper Seward Glacier is subjected to annual freezing and thawing for about one decade.

Localization of ice layers and lenses in the upper and lowermost parts of the 1948–1949 firn layer probably means that the snow of fall and spring has more and thicker thaw crusts than winter snow has. These impervious crusts are particularly favorable for the formation of ice layers.

The association of horizontal ice masses with ice glands is the result of meltwater's spreading horizontally outward from these vertical conduits. The flat bottoms and undulatory upper surfaces of ice layers and lenses are best explained by differential upward growth from an originally flat, impervious horizon. Differential growth could be caused by unequal water supply or by differences in pore size, allowing one part of the firn to retain water more easily and in greater quantity than another. The sharp lower boundaries and ill-defined upper limits of the ice masses presumably reflect the impermeability of the underlying horizon and unconfined upward growth. Continuity of layering between ice masses and the adjoining firn is to be

expected if the ice bodies are formed by freezing of water-saturated firn. The offset arrangement of ice lenses in vertical sections suggests that meltwater moves horizontally along an impermeable horizon until it finds a way to greater depth. Flow of meltwater beyond the ends of ice lenses or through permeable spots in ice layers is essential to the growth of ice bodies at greater depth. Various relations indicate that this has occurred, and it is confirmed by experiments (Ahlmann, 1936, p. 65). All the features described are compatible with the development and growth of these horizontal ice bodies *in situ* within the firn. Most of them are difficult to reconcile with a surface origin and subsequent burial.

In Spitsbergen the tops of annual firn layers are said to be marked by ice layers (Ahlmann, 1935*a*, pp. 32–33, 35); but on upper Seward Glacier they are distinguished by faint dirty discoloration and by firn of coarse grain size and greater porosity. Rarely is ice found along these horizons, and then only in the form of short lenses. Conditions which presumably produced an impermeable crust at the top of annual firn layers in Spitsbergen do not seem to have been attained on Seward Glacier.

The formation and growth of ice masses in firn are diagenetic changes contributing to the conversion of snow and firn to glacier ice. They are not the only processes and cannot act below a depth of about 50 feet in the Seward firn. An estimated 20–25 per cent of the firn is converted to solid ice in this area by meltwater freezing.

ICE GLANDS

Description.—Crudely cylindrical, vertical columns of ice in firn have been observed in Spitsbergen, the Alps, and Alaska. They have been called "ice glands" (Ahlmann, 1936, p. 48), "pipes" (Glen, 1941, p. 143), "firn pipes" (Sharp, 1949, p. 180), and have been described without any name (Moss, 1938, p. 219). Ahlmann (1935*a*, p. 35) likens them to "gnarled trunks," and Seligman (letter, January 19, 1950) has called them "firn trunks." Because they branch horizontally into ice layers and lenses, the analogy with a tree trunk is reasonable, although "ice column" (F. B. Leighton, unpublished manuscript) is suitably descriptive and has the virtue of indicating the composition of the body. In view of the foregoing dispersed usage, Ahlmann's term "ice glands" seems best, as it has priority.

In the Seward firn most ice glands are 4–6 inches in diameter, but many are 2–9 inches, and some attain 2 feet. Lengths are mostly 4–6 feet, and the greatest observed was 8 feet. The glands are roughly cylindrical, strongly ribbed, and have numerous projecting ice layers or lenses (pl. 1, *C*, *D*).

Some glands terminate at both ends in ice layers or lenses, others have a horizontal ice mass only at the lower end, and still others show no relation to horizontal masses at either end. A few inches of horizontal offsetting along ice layers has been observed (fig. 9, *C*), and bifurcation occurs in some instances. A few glands display a slightly tabular form or have dikelike projections. Many of these features are illustrated in sketches (fig. 9) and photographs (pl. 1, *C*, *D*).

Ice glands are composed principally of bluish ice of greater density and larger equidimensional crystals, 5–10 mm. in diameter, than the bulk of the firn. According to Ahlmann (1935*a*, p. 35), they consist of "iced firn," that is, a mass composed partly of solid ice and partly of granular firn. In the Seward area, occasional thin horizontal layers of firn

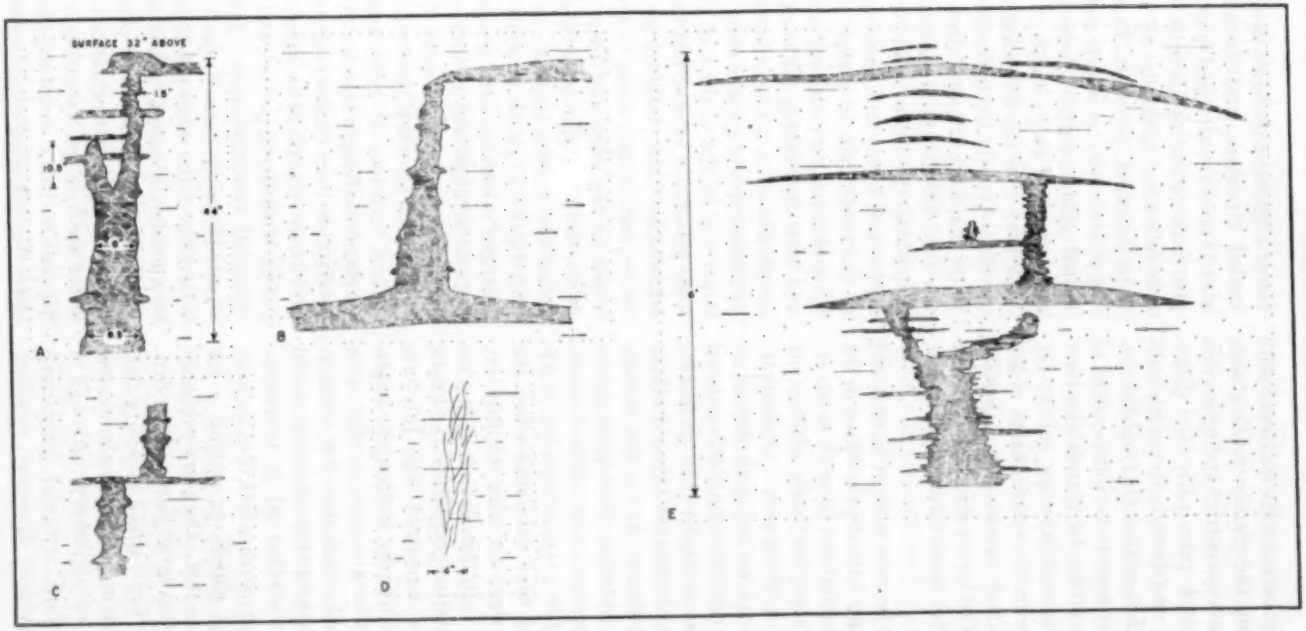


FIG. 9.—Sketches of shapes and relations of ice glands in Seward firn

form partings between the masses of solid ice composing the glands. Some glands display a faint concentric banding in cross section, caused principally by differences in size and distribution of air bubbles. In a thin-section this bubble distribution was seen to be independent of crystal boundaries. Ice glands also have marked horizontal layering (pl. 1, *D*), continuous with the stratification of adjacent firn and presumably inherited therefrom. The layers are 0.5–1.5 inches thick and differ in air content, grain size, and density.

Ice glands become more numerous and generally larger with depth. Those in the uppermost annual firn layer show no accordance in tops, and none reaches the surface early in the season. For example, on June 15, 1949, no ice glands showed on the surface, but they were encountered in excavations at depths of 30–35 inches, and subsequent ablation exposed others that were shallower. As the firn surface is lowered by ablation, the tops of glands appear as low mounds or knobs of coarse ice crystals, 2 or 3 inches high (pl. 2, *A*). Such mounds were numerous when the firn surface was first seen on July 10, 1948, and in 1949 they appeared first on July 6. As ablation progresses, the knobs become more numerous. On July 16, 1948, at the Airstrip Station, 23 knobs were counted in a square, 15 feet on a side. Experience indicates that excavations 3 feet deep in this same area would have exposed another 20 to 25 glands. This area of 225 square feet thus incloses some 40 to 45 glands, or roughly 1 for every 5 square feet. They are probably not so numerous in all parts of the firn.

Where the firn surface is nearly flat, the glands show no particular planimetric arrangement or alignment, although a tendency toward irregular gregarious clustering is evident. However, excellent linear alignment is shown along traces of buried crevasses, and on some slopes gland knobs are arranged in parallel sinuous lines extending directly downslope and following faithfully all changes in slope direction (pl. 2, *B*). These lines on slopes are 2–10 feet apart, and the individual glands within them are spaced at intervals of 10–30 inches.

Origin.—For the following reasons ice glands are attributed to freezing of meltwater that has percolated downward into the firn: (1) They are composed predominantly of ice or iced firn; (2) ice layers and lenses are abnormally thick at the lower ends of glands (fig. 6, *F*); (3) numerous ice lenses project laterally and with diminishing thickness from the glands (fig. 9; pl. 1, *C*), and all horizontal ice masses are of greater abundance and larger size in the vicinity of glands; (4) some glands are located just beneath the edges of ice lenses, indicating that water moved horizontally along the tops of the lenses to the edge and then continued downward; (5) glands ending in horizontal ice masses increase in diameter with depth; (6) glands on some slopes are arranged in sinuous lines similar in spacing and arrangement to meltwater rills, earlier occupying these same slopes.

The freezing of concentrated and channelized invasions of meltwater to form ice glands is thought to occur in spring by conduction from chilled, subzero firn, as previously recognized by Ahlmann (1936, p. 65), Moss (1938, p. 219), and Seligman (personal communication, January 19, 1950). However, before this hypothesis is developed, the other possibility should be considered, namely, that ice glands mark loci of restricted permeability where residual concentrations of meltwater are held for refreezing in the fall. Support for this concept might be drawn from the following experiment. In late

July, 1948, a small area of the Seward firn was liberally sprinkled with fluorescein dye powder and then thinly covered by loose firn, in order to prevent excessive differential ablation by radiation absorption. One month later a 10-foot pit was dug in this dyed area. Most of the dye had been completely flushed through the firn, leaving faint traces of color only in the less pervious parts or in places where meltwater circulation was restricted for other reasons. Some of the colored firn masses had a columnar shape but lacked horizontal projections. The more obvious of these colored columns were located beneath thick parts of ice layers and lenses, supporting the point previously made that horizontal ice masses locally restrict the amount of meltwater moving into the underlying firn.

The formation of ice glands by freezing, in early winter, of meltwater held in impervious or protected masses of firn cannot be easily reconciled with many of the field relations. Furthermore, Sverdrup (1935, pp. 74–75) shows that little water remains in the firn by the time winter chilling sets in. Finally, ice glands formed by June in firn that accumulated during the preceding winter and spring could not possibly have been the result of freezing in the fall.

With regard to the exact mechanics of ice gland formation, two hypotheses are considered: (1) that the glands and associated horizontal ice bodies formed and grew slowly, increment by increment, through freezing of successive invasions of meltwater (this mechanism has some similarity to the development and growth of an icicle) and (2) that the ice glands formed simply by saturation of a columnar mass of firn of the ultimate shape, size, and description which was then frozen. Subsequent growth is not necessarily excluded, but it is not required. The saturated condition was possibly attained in stages, but the freezing, once initiated, was continuous. This mechanism is favored for reasons advanced in the following discussion.

Most firn is more permeable in the thawed than in the frozen state, and meltwater penetration will be deepest, most rapid, and in largest volume in thawed areas. Thawing by radiation penetration does not extend deeper than 20–30 inches. Thawing by conduction from the atmosphere is even more limited, for it ceases, once a thin surficial layer is thawed, as no temperature gradient can exist therein (Sverdrup, 1935, p. 71). Meltwater is the principal means of thawing frozen firn at depth, but, because the water is at 0° C., it can do this only by freezing and liberating the heat of fusion. This would seem to favor the gradual icicle-like development of ice glands; but a simple calculation[4] shows the amount of freezing required to be so small that the only changes resulting are a slight decrease in porosity and a small increase in density of the firn.

During summer on upper Seward Glacier, meltwater is supplied from the surface in a series of daily surges, which drop off sharply or cease completely at night, depending on the weather. This daily fluctuation may be even more marked in spring, but the amount of water supplied then is smaller. At the end of a spring day's thawing, water will

[4] Consider 1 cc. of firn of density 0.50 at −5° C. This material has a porosity of 46 per cent ([0.92−0.50]×100/0.92), assuming the density of pure ice to be 0.92. The specific heat of this firn is approximately 0.25 cal/cc. Raising the original cubic centimeters of firn to 0° C. will require freezing of only 0.0156 gm. of water ([5×0.25]/80), which will increase the firn density to 0.516 and decrease the porosity to 44 per cent. Thus the material remains highly porous, and it is certainly a long way from solid ice.

probably have penetrated the firn to greater depth and in larger volume in some spots than in others because of surface irregularities or initial inhomogeneities. During the succeeding night, some or all of this water may freeze, depending on its amount and the temperature of the firn. Freezing by conduction to the atmosphere occurs only in the uppermost few inches, so freezing at greater depth must be by conduction to the surrounding subzero firn. Early in the season, when firn temperatures are at their lowest, between −10° and −15° C. on upper

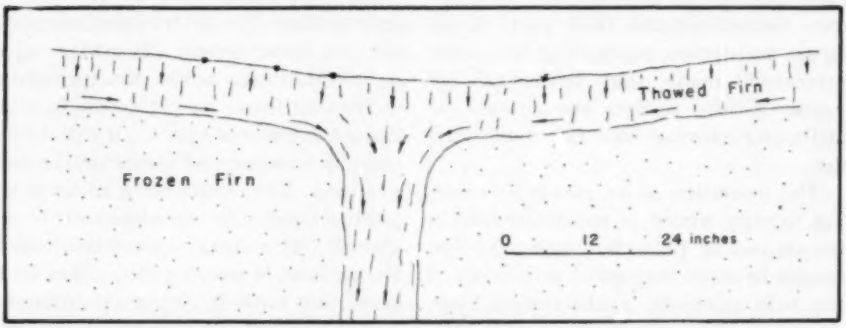


FIG. 10.—Diagrammatic sketch of hypothetical meltwater channelization leading to formation of an ice gland.

Seward Glacier, and the meltwater supply is small, the thawed firn may all be refrozen at night. However, experience with crusts formed at night on thawed firn by conduction to the atmosphere, under temperature differences about comparable to those between frozen and thawed firn, indicates that freezing by conduction is so slow that areas of thawed firn in depth can remain unfrozen for many hours and possibly even for days without further inflow of meltwater. The next meltwater charge then penetrates to greater depth and in greater volume in the unfrozen parts of the firn, owing to their greater permeability (fig. 10). If all the water froze each night, the next day's flow would probably find some other part of the firn more permeable, and progressive development of a water-rich condition would not occur.

The firn layers penetrated by a meltwater column hold different amounts of water, as determined by pore sizes or by impervious underlying horizons. Consequently, the layers within a water-rich column undergo different degrees of icification, and the original firn stratification is preserved or even emphasized. The ribs and ice bodies projecting horizontally from the glands are formed, of course, by lateral movement of water outward along especially pervious layers or on top of impervious horizons. Some firn layers or crusts must be essentially impermeable from the beginning, as indicated by horizontal offsetting of ice glands and by their sudden downward termination at ice layers or lenses (fig. 9, *B*, *E*).

The ice layers and lenses projecting from an ice gland are not easily explained by the hypothesis of incremental growth, for it seems that these horizontal projections would prevent meltwater from following the gland to greater depth. Either the horizontal projections would have to form successively from lowest to

highest, or the interior of the gland would have to maintain some degree of permeability. The first is contrary to all expectations, and there is no evidence for believing that it occurs. As to the alternative, it is known that ice bodies in snow need not be completely impermeable (Work, 1948, p. 541); and Leighton and Miller (personal communication) were able to demonstrate permeability in ice glands near the surface, where the crystals were disarticulated by thawing. Nonetheless, it seems highly improbable that a solid ice gland has sufficient permeability to maintain its own growth and that of associated ice layers and lenses at depth.

The geometric relations are more easily explained by having a columnar mass of firn and its various horizontal projections all attain a water-rich condition at the same time and then freeze as a unit by conduction from surrounding subzero firn. A little initial freezing may be required to thaw the firn, making it more permeable. It also appears likely that the attainment of a water-rich condition requires a period of successive or nearly successive days of surface thawing, during which only a little of the meltwater held at depth refreezes at night. The likelihood of limited freezing is suggested by a report of water-logged firn in contact with dry firn having temperatures down to $-4^{\circ}.5$ C. in Iceland (Ahlmann and Thorarinsson, 1939, pp. 46–47). Any freezing that does occur will be on the periphery of the water-rich area and may develop something of an impermeable jacket enclosing and channelizing further influxes of meltwater.

Some means must be found for shutting off subsequent flows of meltwater so that the water-rich firn can freeze. This control could be climatic; for, in spring, periods of surface thawing during a number of successive days must be succeeded by intervals of similar duration without thawing. Or the locus of meltwater channelization, to be discussed shortly, may change, diverting the flow into other parts of the firn. Before freezing, the meltwater presumably distributes itself within the various firn layers according to the size and continuity of the pores and with relation to underlying impermeable horizons, so that different degrees of icification result. Because the ice glands and associated horizontal ice masses are not wholly impermeable, minor amounts of meltwater may subsequently move into these bodies and produce further icification.

Wherever meltwater moving downward encounters an impermeable horizon, it moves laterally until a path to greater depth is located or until the supply of water is exhausted. This accounts for the termination or offset of an ice gland on an ice layer or lens, for the greater thickness of a horizontal ice mass at its junction with an ice gland, and for the location of a gland beneath the edge of an ice lens. Ice glands extending downward from horizontal ice layers presumably mark the locations of pervious "holes" in those bodies. The discordance in level of the tops of ice glands, not easily explained by the hypothesis of incremental growth, is attributed to the different heights of water-rich columns or to different depths to the top of the winter's chilled layer in the firn when the glands were formed. The fact that ice glands do not initially reach the surface is probably due to the slight degree of chilling in the uppermost snow layers accumulating under the higher air temperatures of late spring and early summer.

If ice glands are formed by freezing of meltwater that has percolated downward from the surface, a satisfactory means of

concentrating the water into columnar channels must be found. Alignment of firn trunks along buried crevasses suggests that meltwater is collected by the topographic and structural sag in firn bridging the crevasse. Similarly, the alignment of glands down a slope suggests that they must be related to small meltwater rills developed on these same slopes earlier in the season. It is not known for certain that a line of glands underlies exactly the course of an earlier meltwater rill, but the aligned glands lie on slopes that were rilled, and they reproduce the trend, alignment, spacing, and adjustment to slope directions of the original rill channels.

These relations suggest that ice glands are indeed related to surface concentrations of meltwater, but why has the downward percolation occurred along columnar channels? The following speculative suggestion is offered for relatively flat areas. When thawing first develops in spring, only the uppermost few inches of the firn are thawed. Meltwater moves laterally through this thawed layer more readily than down into the underlying frozen firn. Consequently, water tends to collect in low spots and, not having anywhere else to go, gradually percolates downward into the firn (fig. 10). A columnar mass of firn underlying a minor surface depression is thus made water-rich. Sverdrup (1935, p. 79) attributes differences in firn temperatures recorded at various localities in West Spitsbergen to irregular meltwater percolation. Even the most level parts of the Seward firn have minor irregularities into which meltwater could be concentrated, and the irregular gregarious grouping of glands beneath relatively flat surfaces suggests that meltwater is concentrated into broad shallow basins, from the floors of which a number of downward-projecting columnar channels are formed. Similar columnar channels could develop in the sags over crevasses, but it is difficult to imagine how or why they form beneath rill channels on slopes. Probably, columnar channelization is a self-enforcing process which gathers strength rapidly, once it is initiated by some minor inhomogeneity within the firn.

In recapitulation, it is postulated that ice glands are formed in the Seward firn in spring and early summer, when the winter's chilled layer is being ameliorated by the downward percolation and refreezing of meltwater. For reasons not clearly understood, much of this meltwater is concentrated into vertical columnar channels. These columnar masses and associated horizontal layers of firn are brought to a water-rich condition by repeated invasions of meltwater on successive or near-successive days. Eventually, the meltwater supply is cut off or diverted and the water-rich mass is frozen by conduction from the surrounding subzero firn. Thus the column of icified firn known as an "ice gland" and its associated horizontally projecting ice lenses and layers are formed. These vertical columns of ice and the associated horizontal masses give the firn much greater structural strength than it would otherwise possess. Ice glands can be expected in firn lying within climatic environments which develop a chilled layer in winter and furnish copious meltwater in late spring and summer. They are probably more widespread than previous reports suggest. Some features of ice glands, such as the concentric banding and the crystal fabric, await thin-section study before explanations can be attempted.

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GEOLOGICAL NOTES

STUDIES OF THE TAKU GLACIER, ALASKA[1]

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Taku Glacier is a relatively large glacier, several hundred square miles in area, which drains southward from the Coast Range into the head of Taku Inlet, northeast of Juneau. Its length, measured to the uppermost part of the névé, is about 34 miles, of which the lower 10 miles varies in width from 1½ to 2 miles. The area of accumulation ranges in elevation from about 3,000 to 6,000 feet, distributed among broad upland névé basins and small tributary glaciers flowing from the surrounding mountains. The terminus until recently was tidal but is now largely separated from the inlet by a frontal moraine, with which the glacier is in contact. Depth measurements indicate that ice near the terminus extends to more than 250 feet below sea-level.

Variations in the position of the terminus have been recorded since the 1890's, but until 1948, so far as is known, no detailed glaciological investigations have been made at higher levels. In that year a series of observations and measurements was begun as part of the program of the American Geographical Society's Juneau Ice Field Research Project, which, since 1949, has been performed under contract with the Office of Naval Research and with the active co-operation of the departments of the Air Force and Army, and of various other government and private agencies. Field parties operated in the summers of 1948, 1949, and 1950, and for one month during the winter of 1951 (Miller, 1949; Poulter, Allen, and Miller, 1949; Field and Miller, 1950; Lawrence, 1950; Bader, 1951; Leighton, 1951). The project is a continuing one, and plans have been drawn up to have a party in the field during the summer of 1951 and at periodic intervals in future years.

Observations are being made to determine the rates of accumulation and ablation, structural features of the névé and underlying ice, the volume transfer of ice from the area of accumulation to the terminus, and the meteorological factors that affect the glacier's regimen. It is hoped that sufficient data will soon be available from which to determine the hydrological balance of the glacier and the main features of its nourishment, flow, and wastage. To date, most of the studies have been carried out near the site of a small research station built in 1949 on a nunatak 16 miles above the terminus and adjacent to the lower part of the area of accumulation. Supplementary observations have been made from a network of subsidiary camps, both up-glacier and toward the terminus, as well as at upper levels along several of the tributary branches. Ground control is being extended toward the heads of major tributaries to the glacier, so that a topographic map can be compiled from the aerial photography which already exists. Depth measurements have also been made by seismic means along four profiles in the lower 16 miles of the glacier for a determination of the volume of ice and the subglacial topography.

Efforts have been begun to determine both surface and englacial movement by means of a vertical profile within the glacier. In 1950 several holes were bored with a rotary core drill at a point where seismic measurements in 1949 had shown the ice to be ap-

[1] Manuscript received March 19, 1951.

proximately 900 feet thick. An aluminum pipe was inserted to a depth of 245 feet in one of the drill holes and is being measured periodically by means of an azimuth inclinometer to determine differential displacement. In addition, a 160-foot length of direct-reading thermistor elements was inserted in a second drill hole for englacial temperature recording at different seasons of the year. Further efforts will be made in 1951 to obtain additional thermal data relating to the depth of penetration of the surface winter cold wave, and at greater depths "boring" will be attempted by means of a "hot-point" unit. Additional geophysical studies are envisioned this year. These third-dimensional measurements and observations, however, remain a most difficult problem to resolve.

Of special interest is the fact that the terminus of Taku Glacier has advanced about 3½ miles since 1900, so that we are dealing with a glacier which may, unlike most others on the continent, have a positive regimen at the present time or at least one close to a state of equilibrium. Regimen studies should be somewhat more conclusive in a glacier that is still maintaining itself and thus is a product of prevailing external causative factors. The Taku may therefore provide a more adequate basis for determining the factors which produce glacier variations and their interrelationships than a glacier experiencing a period of negative economy. Comparative observations, however, should be made with the regimen of neighboring glaciers, whether shrinking or advancing.

Other groups are invited to participate in this long-term study and to develop different lines of inquiry. The area is well adapted for small parties working independently and using as bases the existing facilities of the research station and the outlying camps. The area is accessible by foot, and field parties at the main base camp have been easily supplied during the last three seasons by ski-equipped aircraft operating from the Juneau airport.

It would be highly desirable to institute this type of study on glaciers in other areas where the regimen and relationships to meteorological conditions may also be determined. Similar studies on suitable glaciers (such as could be found in the Cascades and in southern Canada) would be advantageous from various points of view; and direct comparison could then be made with the results being assembled in the two prototype programs of the Juneau Ice Field Research Project and the Arctic Institute of North America's Project "Snow Cornice" in the St. Elias Mountains. These projects, we may hope, represent only the beginning of a more widespread attention to comprehensive glacio-meteorological observations in the Western Hemisphere.

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REVIEWS

Applied Sedimentation. Edited by PARKER D. TRASK. Prepared under direction of the Committee on Symposium on Sedimentation, National Research Council. New York: John Wiley & Sons, Inc., 1950. Pp. v+707. $5.50.

This symposium is "designed (1) to describe aspects of mutual interest to the geologist and to the engineer so that each can understand the other's problems and thus cooperate more effectively in their work; (2) provide information for the consulting geologist who may not be completely familiar with specific problems; and (3) acquaint students with the many practical applications of sedimentation so that they may be more fully informed as to possibilities for a career in this field." To a considerable extent, the book achieves these objectives. In addition, it provides a valuable sourcebook for teachers of geology and engineering and writers of geological and geological engineering textbooks.

The 35 papers in the symposium are written by experts in their respective fields. The contributions are grouped under seven headings: basic principles of sedimentation (6), engineering problems involving strength of sediments (6), applications of processes of sedimentation (11), applications involving nature of constituents (3), economic mineral deposits (4), petroleum geology problems (3), and military applications (2). Admittedly, the coverage of the subject is not complete and obviously cannot be in a single volume. Much of the information is not new, but the book performs a valuable service in bringing together material from widely scattered sources, many of which are unknown or not available to most geologists and engineers. The bibliographies at the end of each paper will prove very useful.

The quality of the papers and coverage of subject matter is good in general but varies from excellent to poor. Particularly regrettable is the inadequacy of the introductory paper, which attempts to summarize the principal features of sediments, the processes by which they form, and the changes that occur subsequent to their deposition. As might be expected, there are some duplication and overlap between papers.

The relation between some of the papers and "sedimentation," as the term is commonly used in geology or engineering, is nebulous. Many of the papers deal with other processes or with the engineering properties of sedimentary rocks. Thus the title *Applied Sedimentation* does not describe the content of the book adequately. It might equally well be called "Engineering Geology." In fact, in many ways the book is a better text for engineering geology than any with which the reviewer is familiar. Certainly, it is a "must" for the library of any teacher of geological engineering or economic geology.

The section on petroleum geology has been restricted to three papers on "Subsurface Techniques," "Porosity, Permeability, and Capillary Properties of Petroleum Reservoirs," and "Carbonate Porosity and Permeability." In large part, these papers are summaries of information available to petroleum geologists, but they should be valuable to other geologists and engineers who are unfamiliar with this field.

The two chapters on military applications, particularly the excellent paper on "Sedimentary Materials in Military Geology," should aid both geologists and engineers in appreciating the importance and value of geology in warfare.

Applied Sedimentation is a contribution to geology and engineering that does credit to the Committee on Symposium on Sedimentation and to the writers who contributed to it. It is a worth-while addition to the library of those interested in applied geology, sedimentation or sedimentary rocks, and to all teachers of elementary geology.

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Minerals and How To Study Them. By EDWARD SALISBURY DANA. 3d ed. Revised by CORNELIUS S. HURLBUT, JR. New York: John Wiley & Sons, Inc., 1949. Pp. x+233; figs. 387; 1 color plate. $3.90.

The first edition of this book was published in 1895 and, as noted in the Preface to the present edition, was "the first of its kind to present facts about minerals in a manner interesting to

the amateur." This is apparently the book's first major revision. According to the Preface, the chapters "Crystals and Crystal Habits" and "Physical Properties of Minerals" cover much the same ground as in the previous editions, but "Chemical Properties of Minerals" and "The Use of the Blowpipe" have been somewhat shortened. The chapter on "The Description of Mineral Species" has been completely rearranged and rewritten: the minerals that are compounds of the same metallic element were grouped together in the first two editions; they are now arranged according to the chemical classification, i.e., native elements, sulfides, oxides, etc.

The volume greatly resembles the much better-known *Dana's Manual of Mineralogy*, widely used as a text in elementary mineralogy and also revised (15th ed., 1941) by Hurlbut. That book is somewhat longer (480 pp.), approximately the same in price ($4.00), and intended for "the student of mineralogy, the mining engineer, the geologist, and the amateur mineralogist." Two questions which may be asked about the present volume are: "What value does it have for the amateur mineralogist that the *Manual* does not have?" and "Might the present volume make a better text than the *Manual* for certain beginning courses in mineralogy?" With these questions in mind, the two books are compared.

The first chapter, "Minerals and Mineralogy" (6 pp.), covers much the same material as the first part of the Introduction to the *Manual*, but in a much more complete and interesting manner. It leads up to the same definition of a mineral, after a discussion of the occurrence of minerals, in a way that would not be out of place in the *Manual*. Chapter ii (6 pp.), "Some Preliminary Hints on How To Study Minerals," discusses the general use of physical and chemical properties in mineral determination. This is much the same material as in the second part of the *Manual*'s Introduction, but is of more interest to the beginner, and ends with a 3-page section on how to collect minerals.

The first 10 pages of chapter iii (56 pp.), "Crystals and Crystal Habits," are much simpler and more general than the corresponding 20 pages of the *Manual*. The definition of a crystal and of its various properties is given. The plane, axis, and center of symmetry are defined, and the measurement of interfacial angles is discussed, but X-ray diffraction is barely mentioned. Following this introductory section, each crystal system is treated in much the same manner as in the *Manual*, but without the complexities of symmetry notation, axial ratio, parameters, and indices, which are omitted also from the preceding section. Instead of a primary breakdown into classes, the various crystal forms are discussed with only a mention to the symmetry class to which each belongs. Essentially the same forms are described and the same drawings used as in the *Manual*. An especially valuable part of chapter iii, developed beyond that in the *Manual*, is the section on malformed crystals, that confusing bugbear of the beginning mineralogy student or collector. The section on twin crystals, more elementary than in the *Manual*, contains descriptions of the chief forms of twins, with a less detailed discussion of twinning laws. The section on crystal habit and crystalline aggregates is expanded into a text version rather than given in outline form, as in the *Manual*. An additional feature of the present edition, and one not found in the *Manual*, is a 5-page section on the growing of crystals.

Chapter iv (23 pp.), "Physical Properties of Minerals," is almost identical with the corresponding chapter, "Physical Mineralogy," in the *Manual*. Chapters v (8 pp.) and vi (20 pp.), "Chemical Properties of Minerals" and "The Use of the Blowpipe," together correspond to the chapter "Chemical Mineralogy" in the *Manual*, to which has been added a brief discussion of chemical elements, valence, and compounds. The section on the blowpipe contains much the same material as the *Manual*, although in a slightly more elementary fashion. It omits entirely the *Manual's* systematic lists of tests for the elements and also the outline entitled "Order of Tests in Blowpipe Analysis."

Next follows a 165-page chapter, "Description of the Mineral Species," corresponding to the 280-page chapter in the *Manual*. There are formal descriptions of 155 mineral species, 43 less than in the *Manual*. Representative of those omitted are stannite, bournonite, collophanite, chalcanthite, arfvedsonite, and phenacite. For each mineral, habit, physical properties, composition, and occurrence are considered. Drawings of the most common crystal forms for most minerals are included; these are the same sketches as appear in the *Manual*, and the descriptions of the minerals are virtually the same. The *Manual* completes this chapter with a 22-page section on "Occurrence and Association of Minerals," which includes a description

of the three classes of rocks and some of their important representatives; this is not included in the volume under review, nor is the 25-page section on "Mineral Uses."

Chapter viii (22 pp.), "On the Determination of Minerals," contains preliminary discussion and an introduction to the 17 pages of determinative tables. In general arrangement these are about the same as those in the *Manual;* only one table is missing, that covering nonmetallic minerals with definitely colored streak. In most cases these are included in the tables covering minerals with metallic or submetallic luster. The column headings are the same, except that composition and crystal systems are not included, and the entry under the "Remarks" column does not contain references to flame tests, etc.

Appendix I consists of a list of the common minerals arranged according to the prominent (metallic) elements they contain, and Appendix II is a list of 102 most important minerals for a small collection. The latter is almost identical with the one in the *Manual*, omitting 15 from that list but containing 5 not on it.

It appears to this reviewer that the present book is well suited to the needs of the young amateur mineralogist or collector who is serious about his mineral collecting and already quite interested in it; one who already owns or has studied such books as English's *Getting Acquainted with Minerals* or Loomis' *Field Book of Common Rocks and Minerals*. It is probably as good as, or preferable to, the *Manual* for a beginning course in mineralogy which is not intended for pre-professional geologists and in which many of the students will not be taking more advanced courses in mineralogy. It is particularly good for those classes in which the training in chemistry of many of the students is weak or nonexistent. However, the *Manual* remains a better text for beginning pre-professional courses in mineralogy and is a better buy for the serious adult amateur mineral collector or student of mineralogy.

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An Index of Mineral Species and Varieties Arranged Chemically. By MAX H. HEY. London: British Museum, 1950. Pp. xx+609; 1 folding chart. £1 10s.

The first half of this book lists minerals according to a chemical classification containing thirty-five major subdivisions. Each mineral is given a number consisting of three parts. Thus under 19.7.12, where 19 means a phosphate, 7 one with Al alone (metals are covered in order of atomic number), we find the twelfth mineral listed is wavellite. Along with the name is given the formula (unit-cell contents, if known), a reference to the literature, and in some cases a few other data.

The last half of the work is an alphabetical listing of some twelve thousand names, together with a reference (to Dana or Hintze in many cases) and the number.

Mineralogists are in debt to Dr. Hey and the trustees of the British Museum for this catalogue; it is indispensable to any laboratory in which modern mineralogical research is being prosecuted; it is also a necessary reference work for libraries.

D. J. FISHER
University of Chicago

Earth Waves. By L. DON LEET. ("Harvard Mon. in Applied Sci.," no. 2.) Cambridge: Harvard University Press, 1950. Pp. 122; figs. 58. $3.00.

This is an excellent little book on exactly the subject expressed by the title. All waves known to be transmitted through the earth are considered without limitation to some one origin, e.g., earthquakes. A particularly valuable feature is the simultaneous consideration and comparison of the waves recorded in geophysical prospecting, large blasts, and earthquakes.

The first chapter, on measurement, also contains a good elementary explanation of wave motion. The explanation of the instruments is clear, simple, and essentially complete. I regard the introduction and use of "gravitational" units as regrettable, but all relevant formulas can easily be changed to an absolute system, and the units play no essential part in understanding the processes involved.

The second chapter, on observed types of waves, is simple and complete. Excellent reproductions of actual seismograms illustrate the identification of the various waves. Even more illustrative material could, in my opinion, have been usefully included here, but later sections contain many more reproductions.

The third chapter, on transmission, constitutes nearly half the book. A very clear, simple explanation of wave paths in earthquakes and blasts is given. A large number of path diagrams,

t-x graphs, and seismograms are reproduced. A very interesting and valuable example is a 40,000-foot reversed refraction profile for which the geologic section, *t-x* graphs, and ten seismograms are given. The complete set of thirty-two seismograms would probably have been prohibitive to reproduce. It would, however, in my opinion, have been useful to have included the four required to illustrate the first 5,000 feet of each profile.

The final chapter contains a brief and informative discussion of microseisms. Anyone desiring a good understanding of the transmission of waves in the earth and the use of such waves in deriving the structure of the earth from earthquakes and blasts can obtain it from this book. The level of physics and mathematics required does not go beyond good Sophomore college courses. On the other hand, the wealth of illustrative material will make it a valuable book even for specialists in the field.

DARRELL S. HUGHES
University of Texas

Das Klima der Vorzeit: Eine Einführung in die Paläoklimatologie. By MARTIN SCHWARZBACH. Stuttgart: Ferdinand Enke Verlag, 1950. Pp. viii+211; figs. 70; tables 25.

This is a handbook on the broad subject of paleoclimatology, written by a geologist who is a professor at the University of Cologne. It begins with the history of the subject in which the Ice Age problem has been particularly prominent. Then follows a condensed description of twenty-five factors and conditions of the modern climate which had a bearing on the climate of the past or which are useful in its interpretation. Next the evidence for past climatic conditions is classified as biological, lithogenetic, and morphological (pp. 16–19). Plants and animals indicate past climatic conditions by the requirements of their closest modern relatives (palms in the Tertiary, reindeer and *Dryas* in the Pleistocene) and by size, color, reef-building, annual rings, etc. Among lithogenetic climatic evidences are different weathering processes, mineralogic-petrographic composition of deposits (limestones, salts, till), and special conditions of sedimentation (varved sediments, loess). Morphological evidence includes stream terraces, frost-action phenomena, cirques, and eskers. The greater part of this chapter presents the evidence for past temperatures (pp. 20–54), precipitation (pp. 54–63), and atmospheric pressure (pp. 63–71) and the annual cycle and long-range climatic changes (pp. 71–84). Succeeding chapters deal with geological time as based mainly on varves (pp. 87–91), with the history of the climate through the ages (pp. 92–161), and with the causes of climatic changes and the numerous hypotheses of explanation (pp. 161–191).

Schwarzbach concludes that the great climatic changes were extremely complex and the result of many collaborating factors. The principal agents are held to have been changes in the radiation of the sun and in the face of the earth (p. 190). Extensive glaciations arose, Schwarzbach believes, when lowered solar radiation *coincided* with favorable outlines of continents and oceans and with high mountain ranges. The fluctuations between the Pleistocene glacials and interglacials are attributed to primary changes in the solar radiation, but climatic variations of several lesser orders to numerous subordinate factors.

The Permian Ice Age, in distinction from the Pleistocene, is believed by Schwarzbach to have required a somewhat different position of the continents. The long ages of warm and particularly dry climates are, after C. E. P. Brooks, explained by southerly winds, which, when the Arctic Ocean was ice-free, brought warm air and warm ocean water to high latitudes and caused storm-free, hot and dry summer weather in the middle latitudes (p. 191).

In a treatise as comprehensive as *Das Klima der Vorzeit*, specialists will naturally find causes for disagreement regarding interpretation and emphasis. And, although the favored hypotheses of climatic changes are attractive because they presume collaboration of several factors, the importance of invoked and rejected possible agents is still unsettled. However, the volume is a storehouse of up-to-date facts which are easily located, thanks to logical arrangement, numerous subheadings, and a good index. The discussions are open-minded. The style is brief and to the point. The reference list is long. In short, the book is useful and excellent.

ERNST ANTEVS
Globe, Arizona

Das Polarisationsmikroskop. By CONRAD BURRI. Basel: Verlag Birkhäuser, 1950. Pp. 296; figs. 168; pls. 4. Fr. 28.80.

Considered as a textbook, this work is rather comprehensive; it is quite comparable to Jo-

hannsen's *Manual*. It contains a 53-page section consisting of two equal parts, one on light in general and one on light in crystals (the indicatrix); this is followed by 50 pages on the microscope and measurements with this tool. Then 52 pages are given to orthoscopic observations, and 48 to conoscopic. The immersion technique is treated in 38 pages, and the universal stage has 31. The book is concluded by an 11-page section on the calculation of extinction angles and by an index that appears to be rather complete (*ca.* 800 items).

The author assumes that the student is familiar with crystal morphology and the stereographic projection. He carries theory through elliptically polarized light for those who have special interests, although such "nonpractical" subjects as conical refraction are not covered. Burri emphasizes the immersion, rather than the thin-section, method; the text is thus suitable for mineralogists, chemists, biologists, etc., and is not restricted to the study of thin-sections (the *petrographic* microscope).

The work is divided into nine chapters designated by letters *A* through *J*. The formulae are numbered A 1, A 2, etc. Thus where one finds a reference to (A 35) on page 173, one must consult the Table of Contents to find what pages are embraced by A and then search these. This trouble could have been avoided if at the top of each left-hand page the publisher had added the proper letter before the chapter heading.

The arrangement is very logical from the point of view of an expert. But the teacher must introduce the student to the subject by considerable hopping around. In general, biaxial phenomena are covered before uniaxial. The immersion technique given near the end of the work should be used on the first day in the laboratory. It seems peculiar that the Christiansen effect is not discussed; instead, Burri suggests (p. 233) the use of monochromatic light when the index of the grain is close to that of the immersion medium.

The book contains a wealth of detailed information for the careful student and is highly recommended to him. References to the current literature are quite common. The work is well written and well printed, attractive in appearance, and remarkably free from errors. Unfortunately, "mμ" is written "$\mu\mu$."

D. J. FISHER

University of Chicago

Hydrology: The Fundamental Basis of Hydraulic Engineering. By DANIEL W. MEAD. 2d ed., rev. New York: McGraw-Hill Book Co., Inc. Pp. xvii+728; figs. 397. $7.50.

The first edition of Mead's *Hydrology* appeared in 1919 at a time when hydrology was almost entirely an empirical science. The author presented, at that time, invaluable observations and conclusions based upon the long and successful practice of his profession and making best use of the available scanty data and limited tools of theoretical analysis. Since then, however, accurate data covering all types of hydrologic phenomena have been collected from greatly expanded data-gathering networks, and most of the methods of theoretical analysis now in use have been developed. Unfortunately, the second edition does not fully reflect this new information.

The reader is impressed by the fact that most of the data, examples, and references included in the second edition are taken from the nineteenth century and the first two decades of the present century. Undoubtedly, most of this information is still valid, but greater emphasis upon the more accurate and up-to-date data, examples, and references now available would lead the reader to a higher degree of confidence.

An introduction to meteorology, presented as discussions of the atmosphere, winds and storms, and atmospheric moisture, in the first few chapters of the book is vital to a full understanding of many types of hydrologic phenomena. The vast field of meteorologic information has been well screened by the author to select for discussion those fundamental items most significant to the hydrologist.

The section on evaporation is disappointing because, from a theoretical viewpoint, it does not explain the recent and well-established theories of turbulent exchange and heat budget and, from a practical viewpoint, it does not describe procedures by which the engineer can estimate amounts of water loss from a reservoir under a given operation schedule, of stream-flow depletion, and of consumptive use of water by crops.

The subject of precipitation is treated exhaustively in chapters vii–xiii. It is notable that the subject of storm transposition and adjustment—of fundamental importance to the flood hydrologist—is not mentioned, whereas the more recent, controversial, and relatively undeveloped subject of artificial production of rain receives some discussion. The section on

droughts is a highly worth-while addition to the book and, from the viewpoint of precipitation, is well presented. The value of the section would be enhanced by a discussion of the effects of droughts upon stream flow. This, in turn, has an important relationship to the selection of critical periods for the design of irrigation, water supply, and navigation projects.

The two chapters on geology, like those on meteorology, are vital to a full understanding of hydrologic phenomena. These chapters deal in a particularly realistic manner with the relationship between knowledge of historical geology and the study of hydrology. A similar treatment of the genesis and morphology of soils would be of equal value to the hydrologist.

The excellence of the two chapters on geology is in sharp and surprising contrast with the highly inadequate treatment of ground water. No reference is made to literature dated later than the year 1916. This completely eliminates any reference to the monumental studies of all phases of ground water by the late Oscar Meinzer.

Chapters xvii–xx present background information dealing with runoff, stream flow, and floods. From the viewpoint of seasonal and annual runoff and of stream regulation to meet various requirements, the treatment is fairly complete. On the other hand, there is no discussion of the stream-flow hydrograph in terms of its various components and practically no discussion of methods of hydrographic analysis and synthesis for the various needs of the flood hydrologist.

The outlines presented in the final chapter of factors to be considered in the planning of various types of water resources developments will serve as useful guides to the student. On the other hand, the book presents few procedures and techniques which can be applied directly without the benefit of other experience or guidance in the solution of practical problems.

HERBERT S. RIESBOL

U.S. Bureau of Reclamation
Denver

A German and English Glossary of Geographical Terms. By ERIC FISCHER and FRANCIS E. ELLIOTT. Edited by ANASTASIA VON BURKALOW. ("Am. Geog. Soc. Library ser.," no. 5.) Cambridge, Mass.: University Press, Inc., 1950. Pp. vii+111. $3.00.

Although the compilers have prepared this glossary expecially for geographers, they have included a large number of terms from the related fields of geology, and particularly from geomorphology. Typical pages contain 40–60 per cent of terms commonly used in geology. This volume has the advantage over a standard dictionary, that one may look up, in addition to many simple and compound words, such phrases as: *aufgestiegen an einer Verwerfung* ("upfaulted"), *Folgefluss zweiter Ordnung* ("resequent stream"), and *kulissenförmige Anordnung von Bergketten* ("mountin ranges arranged en echelon").

The last fifty pages of the book consist of an English-German glossary in which one may find the German expressions for such phrases as "elbow of stream capture," "glacial outwash plain," "knob and kettle topography," and "trellis drainage pattern."

The book is necessarily not complete from the point of view of a geologist, but it seems a valuable supplement to the standard German dictionaries in the libraries of geomorphologists and glacial geologists.

JEAN G. SIMMONS

University of Chicago

California Fossils for the Field Geologist. By HUBERT G. SCHENCK and A. MYRA KEEN. Stanford, Calif.: Stanford University Press, 1950. Pp. 88; pls. 56. $2.00.

First issued privately in 1940, this booklet (spiral-bound, standard field notebook size) has been reissued with little change except for the addition of a bibliography. It contains fifty-six plates, of which the first sixteen deal with morphologic and systematic terminology of various faunal forms. The remaining forty illustrate the stratigraphic sequence of forms in the Upper Jurassic to Pliocene formations of California. The booklet is intended for use in rapid field identification of fossils and attempts nothing more. There are three indexes—systematic, stratigraphic, and alphabetic—and a provisional correlation chart.

JEAN G. SIMMONS

University of Chicago

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